INFLUENCE OF LAKES AND PEATLANDS ON GROUNDWATER CONTRIBUTION TO BOREAL STREAMFLOW

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ABSTRACT

How much groundwater flows to boreal streams depends on the relative contributions from each landscape unit (forested uplands, lakes, and peatlands) within a catchment along with its hydrogeologic setting. Although there is an understanding of the hydrologic processes that regulate groundwater outputs from individual landscape units to their underlying aquifers (both coarse- and fine-textured) in the boreal forest, less understood is how the topography, typology, and topology (i.e. hydrologic connectivity) of the landscape units regulates groundwater flow to streams. Improved understanding of groundwater-stream interactions in the Boreal Plain of Alberta and Saskatchewan is critical as this region is undergoing substantial environmental change from land cover disturbances for energy and forestry industries and climate change. This thesis determines groundwater-stream interactions during the autumn low-flow period in a 97 km^2 glacial outwash sub-catchment of White Gull Creek Research Basin, Boreal Ecosystem Research and Modelling Site, Saskatchewan. The catchment (Pine Fen Creek) is comprised of a large (30 km^2) valley-bottom peatland, two lakes, and jack pine (*Pinus banksiana*) uplands. The pine uplands are important areas of annual groundwater recharge for the catchment. Vertical hydraulic gradients (VHGs) show frequent flow reversals between the lakes and sand aquifer, and spatially diverse VHGs between the peatland and sand aquifer. Groundwater flow nets and lateral hydraulic gradients indicate the stream receives groundwater along its length. Isotopic samples of end members corroborate the hydrometric data. Catchment streamflow response during the 2011 low flow period was not simply the addition of net groundwater inputs from each landscape unit. Instead, the large size, valley-bottom position, and short water 'memory' of the peatland were the critical factors in regulation of catchment streamflow during low flow periods. Peatland hydrologic function alternated between a source and sink of runoff (surface and

subsurface) to the stream, dependent on the position of the water table; a value of 0.15 m below peat surface was the critical functional tipping point. Given the high percentage of peatlands (21%) within the Boreal Plain, incorporating their runoff threshold is required in parameterizing runoff generation in hydrological models, and thus predicting impacts of peatland degradation and forest clearing on streamflow.

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LIST OF ABBREVIATIONS

b	Soil Depth [m]
BERMS	Boreal Ecosystem Research and Modelling Sites
d	Snow Depth [m]
E	Evaporation [mm day^{-1}]
EC	Specific Conductance [μs cm^{-1}]
ET	Evapotranspiration [mm day^{-1}]
FEN	Sandhill Fen; Location of BERMS Eddy Flux Tower
G^{I}_{calc}	Ispuchaw Lake Calculated Net Groundwater Exchange [mm day^{-1}]
G^{M}_{calc}	Mature Pine Upland Calculated Net Groundwater Exchange [mm day^{-1}]
G^{P}_{calc}	Peatland Calculated Net Groundwater Exchange [mm day^{-1}]
G^{R}_{calc}	Regenerating Pine Upland Calculated Net Groundwater Exchange [mm day^{-1}]
G^{Z}_{calc}	Zeden Lake Calculated Net Groundwater Exchange [mm day^{-1}]
G^{I}_{meas}	Ispuchaw Lake Measured Net Groundwater Exchange [mm day^{-1}]
G^{Z}_{meas}	Zeden Lake Measured Net Groundwater Exchange [mm day^{-1}]
G^{M}_{r}	Mature Pine Upland Measured Net Groundwater Recharge [mm day^{-1}]
G^{P}_{r}	Peatland Measured Net Groundwater Recharge [mm day^{-1}]
G^{R}_{r}	Regenerating Pine Upland Measured Net Groundwater Recharge [mm day^{-1}]
G_{outlet}	Lateral Groundwater Flow at Outlet [mm day^{-1}]
G_{peat}	Stable Isotopic Composition of Groundwater in Peat [‰]
$G_{sand(1m)}$	Stable Isotopic Composition of Groundwater in Sand at -1 m [‰]
$G_{sand(7m)}$	Stable Isotopic Composition of Groundwater in Sand at -7 m [‰]
$\bar{H}$	Mean Water Table above Aquifer Base [mm]

H94 ....Harvested Jack Pine in 1994; Location of BERMS Eddy Flux Tower

k_c .... Unit Converter from W m^{-2} d^{-1} to mm d^{-1} [0.353]

K_h ....Saturated Horizontal Hydraulic Conductivity [m s^{-1}]

K_v .... Saturated Vertical Hydraulic Conductivity [m s^{-1}]

L ....Distance Between Well and Pine Fen Creek Stage [mm]

LHGs ....Lateral Hydraulic Gradients (m m^{-1})

n ....Number of Samples [unitless]

OBS ....Old Black Spruce; Location of BERMS Eddy Flux Tower

OJP ....Old Jack Pine; Location of BERMS Eddy Flux Tower

P .... Precipitation [mm day^{-1}]

P_e ....Event Rainfall Volume [mm]

P_i .... Event Rainfall Intensity [mm h^{-1}]

PFCC ....Pine Fen Creek sub-Catchment

Q ....Observed Stream Discharge Before Event [m day^{-1}]

Q_e .... Event Runoff [m day^{-1}]

Q_i ....Total Peatland Surface Inflow [mm day^{-1}]

Q_L ....Ispuchaw Lake Outflow [mm day^{-1}]

Q_o .... Peatland Surface Outflow [mm day^{-1}]

Q_S ....Peatland Surface Inflow from Pine Fen Creek [mm day^{-1}]

Q_T .... Peatland Surface Inflow from Western Peatland Tributary [mm day^{-1}]

Q_t ....Calculated Flow from Recession Curve [m day^{-1}]

Q_v .... Latent Water Vapour Content [W m^{-2}]

Q_{120} ....Peatland Surface Inflow measured at Hwy 120 [mm day^{-1}]

S_y ....Specific Yield [unitless]

SMOW .... Standard Mean Ocean Water

SWE ..... Snow Water Equivalent [kg m^{-2}]

t ..... Time [day]

t* ..... Recession Coefficient [day]

T^3 ..... Topography, Topology and Typology

VHGs ..... Vertical Hydraulic Gradients (m m^{-1})

w ..... Width [m]

WGCRB ..... White Gull Creek Research Basin

x ..... Distance Between Well and Groundwater Divide [mm]

z ..... Depth [m]

z_w ..... Water Table Depth [mm]

$\delta^{18}O$ ..... Stable Isotopic Composition of $^{18}O/^{16}O$ (‰)

$\delta^{2}H$ ..... Stable Isotopic Composition of $^{2}H/^{1}H$ (‰)

Δh ..... Change in Hydraulic Head [m]

Δl ..... Change in Distance [m]

ΔS ..... Change in Storage [mm day^{-1}]

ΔS_s ..... Saturated Storage Change [mm day^{-1}]

ΔS_u ..... Unsaturated Storage Change [mm day^{-1}]

$\Delta\theta$ ..... Daily Change in Soil Moisture Content [unitless]

λ_v ..... Latent Heat of Vapourization of Water [MJ kg^{-1}]

ρ_s ..... Snow Density [kg m^{-3}]

ρ_w ..... Water Density [kg m^{-3}]

CHAPTER 1 - INTRODUCTION

1.1 Introduction

The Boreal Plain ecozone of the Canadian Boreal forest is distinguished from the adjacent Boreal Shield and Cordillera regions by its geology. The mountainous Cordillera and Shield regions have relatively shallow soils overlying bedrock, whereas the Plain has substantial glacial deposits (up to 340 m). These geological differences lead to a varied capacity for groundwater storage and complexity in groundwater flows. Within the Boreal Plain, outwash landscapes are typically connected to larger groundwater flow systems because of their permeable coarse-textured substrate (Tóth, 1963), whereas lower permeable clay till moraine results in slower groundwater movement and local groundwater flow systems (Ferone and Devito, 2004). Groundwater can be a key source of low flows in streams (termed baseflow) during the dry season or in periods of drought in outwash landscapes (Smakhtin, 2001), which is critical to maintaining stream ecological integrity (Poff et al., 1997).

Catchments that provide groundwater to boreal streams represent the relative contributions from a mosaic of landscape units including peatlands, lakes and forested uplands. Each landscape unit is individually capable of 'collecting', 'storing' and 'discharging' water (Black, 1997). Buttle (2006) argued that the hydrological function of a landscape unit can be predicted from its topography, topology, and typology, a concept that he referred to as the T^3 template. Hydrologic processes that control groundwater interaction between landscape units and their underlying aquifer have been intensively studied. Groundwater transmission between landscape units has received recent attention, but the studies published do not represent the full breadth of the landscape unit arrangements observed in the Boreal Plain, nor do they describe how these hydrologic processes interconnect to govern groundwater flow to streams. There is a pressing need to advance the understanding of groundwater-stream connectivity in the Boreal Plain as it is an area experiencing unprecedented industrial development of oil, gas and forest resources (Seitz et al., 2011; Devito et al., 2012) and rapid climate change (Schindler and Donahue, 2006; Bergengren et al., 2011; Flanagan et al., 2011).

1.2 Literature Review

This section provides a general review of Boreal Plain hydrology, focusing on groundwater in the common landscape units. Also discussed is the state of knowledge of hydrological connectivity of landscape units. Research gaps studied in this thesis are then identified.

1.2.1 Boreal Plain

The Boreal Plain ecozone extends across portions of Alberta, Saskatchewan and Manitoba with varying thickness of glacial deposits (<1 to >300 m; Fenton et al., 1994). As a result of its surficial geology, Devito et al. (2005b) argue that the Boreal Plain has some of the most complex groundwater and surface water interactions of any ecosystem in the world. Lakes account for 10-15% of open water surface area (Granger and Hedstrom, 2010) and peatlands represent 21% of the Boreal Plain land base (Vitt et al., 2002). Uplands are comprised of trembling aspen (*Populus tremuloides*), jack pine (*Pinus banksiana*), white (*Picea glauca*) or black (*Picea mariana*) spruce species. These landscape units have been studied on a heterogeneous mix of outwash, moraine and lowland plain glacial landforms in north-central Alberta (e.g. Ferone and Devito, 2004; Smerdon et al., 2005). The interaction between surface water and groundwater has also been studied in other deep glaciated terrains, similar to the Boreal Plain, including Wisconsin (Jacquet, 1976; Watters and Stanley, 2007), Minnesota (Siegel and Glaser 1987; Winter et al, 2001) and Nebraska (Winter, 1986).

1.2.2 Groundwater Flow within Landscape Units

Whether individual landscape units function as sources of or sinks for groundwater will depend on their water balance, which is influenced by climate, vegetation, soil properties, and hydraulic gradients. Described below is the state of knowledge of the hydrological attributes for each of the three landscape units common in the Boreal Plain, and how these influence their capacity to act as a groundwater sink or source.

1.2.2.1 Forested Uplands

Groundwater flow in forested uplands (also known as hillslopes) is largely driven by recharge, which is the amount of atmospheric water that percolates into aquifers. Groundwater recharge is defined by Freeze and Cherry (1979) as "entry into the saturated zone of water made available at the water table surface, together with the associated flow away from the water table within the saturated zone". Recharge rates depend on soil texture, the timing and amount of precipitation and evapotranspiration, antecedent conditions, and forest composition. Results from an irrigation plot study found 95% of water inputs infiltrate the sandy pine uplands of the Alberta Boreal Plain due to greater vertical hydraulic conductivity (Redding, 2009). The increased permeability and low water holding capacity of sandy soils (Saxton et al., 1986) allows water to percolate below the root zone and recharge groundwater at a greater rate than fine-textured soils. Greater recharge in coarse-textured uplands occurs during spring freshet and fall, when plant transpiration is at a minimum (Smerdon et al., 2008). The occurrence of recharge in forested outwash landscapes will depend on year-to-year weather variation, which drives whether the site is experiencing a water deficit, water surplus or near balance conditions, and is correlated with annual snow accumulation (Smerdon et al., 2008). Redding (2009) further showed that the timing and intensity of precipitation in the summer and fall can influence recharge to the groundwater regime and prime the hillslopes to provide recharge conditions in subsequent years.

Forest age also influences groundwater recharge. Elliot et al. (1998) demonstrated that younger trees in regenerating pine stands recharge groundwater at a greater rate compared to mature pine stands due to reduced transpiration and canopy interception, and increased soil moisture at depth.

Groundwater recharge will control the water table configuration, which ultimately influences the direction of groundwater flow from upland to a stream (Winter, 1986). Groundwater flow is typically initiated as precipitation infiltrates the soil and moves vertically to an underlying impeding flow layer where it is redirected as lateral flow above the layer of restricted permeability (McDonnell, 1990; Weiler and McDonnell, 2004). Lateral flow has shown to occur at the soil-bedrock interface in steep terrain and humid environments including the Boreal Shield (Peters et al., 1995; Buttle and McDonald, 2002), mountainous regions of New Zealand (McGlynn et al., 2002) and southeastern USA (Tromp-van Meerveld and McDonnell, 2006). However, in areas with considerable depth to bedrock or permeable bedrock, such as many glacial landscapes, preferential vertical flow can be redirected to lateral flow at the interface of a coarse-textured layer over a finer-textured layer with minimal seepage losses to deep groundwater (Winter et al., 2001; Lin et al., 2005; Devito et al., 2005a; Haynes and Mitchell, 2012). In glaciated terrain, lateral flow is controlled by the transmissivity feedback mechanism (Rodhe, 1987; Redding and Devito, 2010; Haynes and Mitchell, 2012). Vertical recharge into the soil must first occur before water tables rise into the more transmissive mineral soil to initiate lateral flow via the transmissivity feedback mechanism (Rodhe, 1987).

Vertical flow will dominate coarse-textured outwash owing to greater vertical hydraulic conductivity and deeper infiltration depths than fine-textured clay till soils (Fig. 1.1; Devito et al., 2012). This results in gentle water table gradients that mirror the underlying restricted permeability layer rather than surface topography (Smerdon et al., 2005). At the hillslope scale, the rate of groundwater flux will vary spatially because the water table will be closer to the land surface at the toe slope compared to the upland crest (e.g. Smerdon et al., 2008). This will influence how subsurface flow responds to water inputs on the hillslope. For example, large water table responses to rain or snowmelt events are possible at toe slope locations because the capillary fringe extends near to the ground surface (Rosenberry and Winter, 1997; Redding, 2009). As the thickness of the unsaturated zone increases, the time for infiltrating rainfall to reach the water table increases resulting in a smoother hydrograph response (e.g. Cuthbert, 2010).

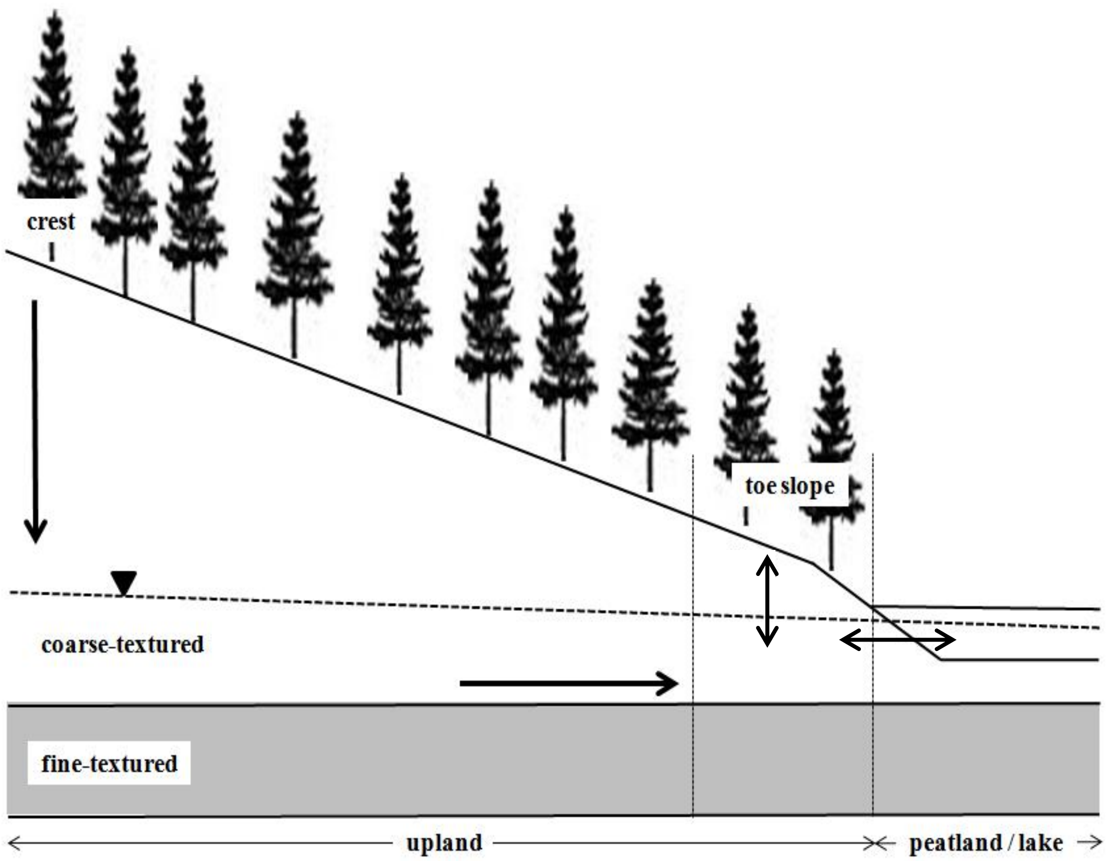


Figure 1.1: Schematic diagram of groundwater recharge and flow in a coarse-textured (e.g. sand) pine upland of the Boreal Plain underlain by fine-textured (e.g. clay-rich till) substrate. After Winter (2001) and Redding (2009).

1.2.2.2 Lakes

Lakes and ponds typically occur where the water table intersects the land surface at the shoreline (Winter, 1986). In a theoretical study, Pfannkuch and Winter (1984) showed that groundwater flux under lakebeds is greatest near the shore and decreases toward the lake center, a result confirmed by several field studies (Attanayake and Waller, 1988; Shaw and Prepas, 1990; Schafran and Driscoll, 1993). The heterogeneity in lakebed sediments can cause variable lake-aquifer hydraulic connectivity resulting in spatially varying groundwater flux (Kishel and Gerla, 2002). However, a greater lake bed surface area would typically result in proportionally more groundwater exchange with the underlying aquifer (Smerdon et al., 2005).

Geologic properties can also affect the relative rate of groundwater flux (Sophocleous, 2002). Lakes located on highly permeable sandy sediments have higher lake-groundwater connectivity, as has been demonstrated in Nebraska (Winter, 1986), Minnesota (LaBaugh et al., 1995; Rosenberry et al., 2000), Wisconsin (Jaquet, 1976; Cherkauer and Zager, 1989; Krabbenhoft et al., 1990), and Alberta's Boreal Plain (Smerdon et al., 2005); however, the magnitude of groundwater flux may decrease with the presence of a low permeability gyttja layer on lake beds (Smerdon et al., 2005). Lakes located on materials with low hydraulic conductivity, such as clay till, will have minimal or no interaction with deeper groundwater (Ferone and Devito, 2004).

Depending on the topography and geology, a lake or pond can function wholly as a source or sink of groundwater recharge to the underlying aquifer or at the same time have distinct source or sink zones (Born et al., 1979), which is influenced by the lakes' position within the groundwater-shed (Cherkauer and Zager, 1989; Smith and Townley, 2002). The direction of groundwater flow may also change seasonally or on shorter time scales. For example, the development of groundwater ridging in the upland due to increased recharge can lead to large and rapid increases in hydraulic head during storm periods or spring freshet, resulting in the lake gaining water from the groundwater system (Anderson and Munter, 1981; Cherkauer and Zager, 1989). However, in summer, several researchers have shown flow reversals wherein the high evapotranspiration demand by upland vegetation removes the water table mound resulting in the pond or lake losing water to the groundwater system (Rosenberry and Winter, 1997; Hayashi et al., 1998; Smerdon et al., 2005).

1.2.2.3 Peatlands

Peatlands are differentiated from other wetlands by an accumulation of at least 0.4 m of peat (National Wetlands Working Group, 1988). Within the Boreal Plain, 36% of peatlands are bogs, 35% treed fens, and 29% open (non-treed) fens (Vitt et al., 2002). Fens receive significant groundwater fluxes, whereas bogs receive the majority of water inputs from precipitation; both bogs and fens can occur within one peatland system (Siegel and Glaser, 1987). Peatlands can be located within valley-bottoms (Hogan et al., 2006), riparian area around lakes or ponds (i.e. pond-peatland complex; Ferone and Devito, 2004) or perched in the upland (Holden, 2009). Peatlands within the boreal forest often have low slope and rough microtopography, resulting in large depression storage capacity (Quinton and Roulet, 1998; Metcalfe and Buttle, 1999). Water storage in peatlands and their contribution to runoff is dependent on climate (e.g. Siegel et al., 1995; Branfireun and Roulet, 1998; Rouse, 2000), hydraulic and thermal properties of peat (e.g. Price and Fitzgibbon, 1987; Hogan et al., 2006; Wright et al., 2009), water table depth (e.g. Verry et al., 1988; Kværner and Kløve, 2008; Jager et al., 2009), and underlying mineral substrate (Siegel and Glaser, 1987; Todd et al., 2006).

A two-layer model has been used to describe peatlands. The upper layer of dead and poorly decomposed vegetation (i.e. fibric to mesic) in the acrotelm has a higher hydraulic conductivity. In the catotelm, the peat is less permeable and is composed of highly decomposed (i.e. humic) plant material (Boelter, 1969). However, Morris et al. (2011b) have indicated that this conceptual model is too simplistic for hydrologic purposes, and instead suggest that the boundary between the acrotelm and catotelm is probably more of a continuous transition (Fig. 1.2).

Peat has a high water content and large compressibility (Price and Schlothauzer, 1999). Changes in peat compression (i.e. vertical displacement) result from water table fluctuation (Price, 2003) and flow processes both seasonally and long-term (Whittington and Price, 2006). Compression affects the main hydraulic properties of peat including its bulk density, hydraulic conductivity, and specific yield (Chason and Siegel, 1986; Hogan et al., 2006). Compression of the peat surface in response to the lowering water table (i.e. dry period) will decrease hydraulic conductivity and specific yield while increasing bulk density (Hogan et al., 2006; Whittington and Price, 2006). The change in hydraulic properties as a result of peat volume change will directly influence transient water storage and groundwater flow through the peatland. For example, water balance analysis of a patterned fen in the Boreal Plain showed large inter-annual variability in groundwater storage between wet and dry years likely due to a combination of changes of the depth of the water table relative to the peat surface and movement of the peat surface due to compression and expansion of the entire peat layer (Barr et al., 2012). The peatland's soil structure functions to conserve water and maintain the water table closer to surface, which results in short-term water 'memory' that fills relatively quickly in response to short term deviations in moisture surplus compared to forested uplands (Devito et al., 2012). Forested uplands typically have long water 'memory' needing several to many years of large moisture surplus to fill the large available storage capacity. Therefore, the return period for runoff from a peatland is approximately one to two years (Devito et al., 2012).

Predicting the spatial and temporal direction of groundwater flow in peatlands is difficult due to the anisotropy and variable storage capacity of peat (Siegel et al., 1995). Researchers have observed that different mechanisms generate and regulate peatland outflow at low-flow (Devito et al., 1997; Hogan et al., 2006; Kværner and Kløve, 2008), compared to rain-driven (Branfireun and Roulet, 1998; Quinton and Roulet, 1998) and snowmelt-driven events (Jager et al., 2009; Spence et al., 2011). Groundwater has been shown to preferentially flow laterally in zones of high horizontal conductivity, specifically in the upper 0.50 m of the peat column (Chason and Siegel, 1986), due to lower vertical saturated hydraulic conductivity in peat (e.g. Fraser et al., 2001; Hogan et al., 2006). Though, horizontal hydraulic conductivity is often quite variable as soil pipes increase water transmission (Holden, 2005). These are found along passages adjacent to major tree roots created by decay of buried logs (Waddington, 2003). Previous research that has focused on flow through the more conductive surface layers of peat illustrates the importance of knowing the position of the water table as it controls the peatland runoff response to precipitation inputs (e.g. Jager et al., 2009). Early work in Minnesota found lateral flow in the peatland increases when the water table is in the less decomposed peat layers or above the peat surface and subsequently influenced streamflow response (Bay, 1969; Verry et al., 1988). Building on that work, water table - discharge relationships have been observed in other boreal peatlands located in the Cordillera (Wright et al., 2009); Shield (Branfireun and Roulet, 1988), Norway (Kværner and Kløve, 2008) and Finland (Jager et al., 2009). Jager et al. (2009) demonstrated that the water table - discharge relationship varies with both annual and seasonal climate patterns (i.e. drier summers, snowmelt runoff) and found near surface lateral peatland discharge stopped when the water table dropped below -13 cm. Others have found that vertical flow increases when the water table drops into the more decomposed peat layers and results in increased interaction with the underlying permeable mineral substrate; as is commonly the situation where sandy substrates underlie peatlands (Siegel and Glaser, 1987; Reeve et al., 2000; Hogan et al., 2006). The dominance of vertical or lateral flow patterns within peat is variable in time and space, and depends on the season, water table depth, and the permeability of the underlying mineral substrate.

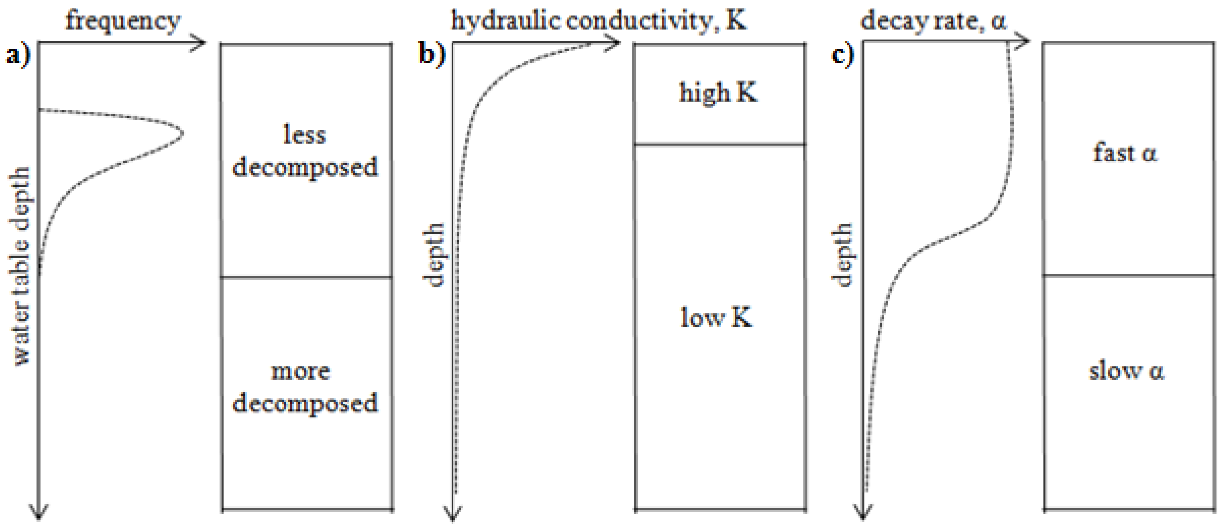


Figure 1.2: Hypothetical peat depth profiles showing: a) frequency distribution of water-table depth in less decomposed (acrotelm, top box) and more decomposed (catotelm, bottom box) layers; b) continuous variation in hydraulic conductivity with depth; and c) continuous variation in peat decay rates. Modified from Morris et al. (2011).

1.2.3 Hydrologic Connectivity and Streamflow Response

Hydrological connectivity, defined by Bracken and Croke (2007), is the ability to transfer water from one part of a landscape to another. Streams receive inputs that are distributed over all or part of their drainage network area. Conceptual models evaluating the hydrologic controls on streamflow production in catchments have been put forward. Devito et al.'s (2005b) hydrologic response unit framework, constructed within the Boreal Plain, uses a hierarchical approach to examine the controls on hydrologic processes that move water through a catchment within a given region using the sequence: climate - bedrock geology - surficial geology - soil depth and type - topography and drainage network. The authors argue this framework would enable the user to define the scale of interaction and create an appropriate conceptual model that could be used to understand the source and flow paths of groundwater flow to streams. Rather than consider each factor controlling water movement within a catchment hierarchically, Buttle's (2006) T^3 conceptual model assesses the interaction of three first order controls: topography, typology, and topology simultaneously. Topography reflects the role hydraulic gradients drive water fluxes from hillslopes to other landscape units (lakes, peatlands, streams), in terms of profile changes in hydraulic gradient and planar (converge versus diverge) form. Typology (e.g. geology, vegetation, soil type, relative size) describes the differential ability of landscape units to store and release water to the stream channel. Topology refers to the spatial arrangement of landscape units and their connectivity, and thus provides insight into their ability to move water to the stream network. The T^3 template can be applied at various spatial scales, but does not directly consider climate.

Understanding the interconnection of landscape units is necessary because groundwater flow to streams reflects the relative contributions from each of the landscape units that make up the drainage network. In the mountains of Montana, Jensco et al. (2009) demonstrated topography (i.e. flow path distance and gradient) and topology (i.e. arrangement of landscape unit structure) controlled the magnitude of runoff between hillslope-riparian area-stream at the catchment scale. Soulsby et al. (2006) showed that Scotland sub-catchments dominated by peat or shallow alpine soils produced smaller groundwater contributions to streamflow compared to freely draining soils; thereby, demonstrating typology was a critical factor in influencing streamflow response. The factors that control the hydrologic connectivity between the landscape units generating runoff and the various flow paths that link landscape units to streams can be difficult to identify due to varying storage, residence time or transit time of water (McGuire and McDonnell, 2006; McDonnell et al., 2007). Therefore, stable isotope analysis of source water has been employed in numerous studies to provide insight into groundwater-stream interaction (e.g. Gibson et al., 2002; Hayashi et al., 2004; McGuire and McDonald, 2010; Spence et al., 2011).

Catchment streamflow response is a function of the connectivity of landscape units to the stream network dependent on their variable storage capacities in the drainage area and along the drainage network (Phillips et al., 2011). Hydrologic connectivity proved important for the non-linear runoff response observed in several catchments with a wide variety of landscape units (forested uplands, peatland, riparian area, wetlands, lakes) including the Boreal Plains (Devito et al., 2005a), subarctic (Spence and Woo, 2003; Spence, 2006; Spence et al., 2011), Precambrian Shield (Branfireun and Roulet, 1998; Buttle et al., 2004), prairie (Shaw et al., 2012), and mountain forests (McGuire et al., 2005; Tromp van Meerveld and McDonnell, 2006; Jensco et al., 2009).

Lakes and peatlands, when located along the core stream channel network (Fig 1.3), have been shown to attenuate the runoff response in both wet and dry conditions as water is routed to the catchment outlet as a result of their threshold storage capacities (e.g. Phillips et al., 2011; Spence et al., 2011). The distribution (e.g. along a watercourse) and topology of lakes within a catchment delay the transfer of runoff from hillslopes or upstream lakes until the lake fills and then spills over its outlet elevation (Spence, 2006; Phillips et al., 2011). Several studies in boreal locations in the subarctic (Spence et al., 2011), Precambrian Shield (Branfireun and Roulet, 1998; Todd et al., 2006), and Norway (Kværner and Kløve, 2008) have shown peatlands located in valley-bottoms regulate the streamflow response. Branfireun and Roulet (1998) suggested streamflow response was controlled by the position of the peatland water table when the catchment was drier, and the antecedent soil moisture conditions in the upland when the catchment was wetter, whereas Kværner and Kløve (2008) found a peatland attenuated groundwater flow differently during low-flow and high-flow (rainfall-runoff events) periods releasing water at different rates as a result of declining hydraulic conductivity with peat depth. The soil structure of a patterned fen (~6 km^2) in the Boreal Plain functioned to conserve water and maintain the water table close to surface during dry years resulting in a shorter runoff-return period with subsequent moisture surplus years (Barr et al., 2012). Thus, the storage capacity of both lakes and peatlands can maintain low flows during dry periods through supplying a steady source of groundwater (Smakhtin, 2001) or act as a sink for groundwater in wet periods (LaBaugh et al., 1997).

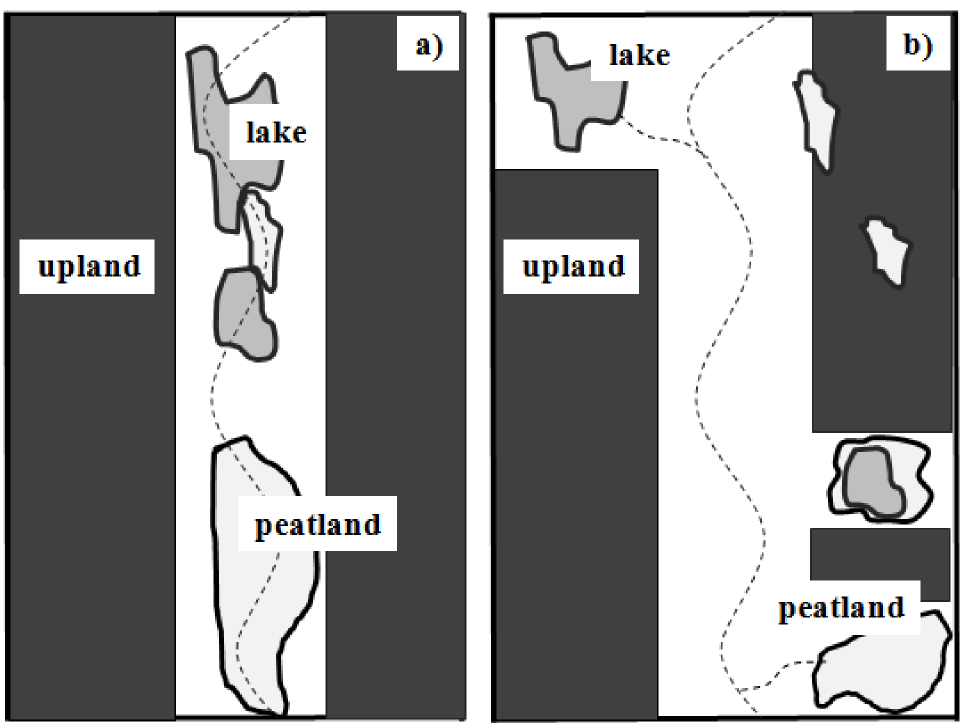


Figure 1.3: Hypothetic catchments with varying spatial arrangements of upland (dark grey), lakes (medium grey) and peatland (light grey) highlighting: a) lakes and peatlands located coincident along the main watercourse (dashed line); b) lakes and peatlands (e.g. perched, pond-peatland complex) located away from the main watercourse. Some landscape units are connected via surface outflow, whereas others are more well-connected to local or large scale groundwater flow systems depending on surficial geology.

Within the Boreal Plain, how groundwater flows through peatlands and lakes as it moves from uplands to streams has not been an area of focus, particularly in outwash sediments. The only exception is a water balance study in the Alberta Boreal Plain underlain by ablation till (Devito et al., 2005a). They reported that large inter-annual variability in soil water deficits and soil water storage capacity of a wetland located in a valley-bottom position influenced the runoff-generation and streamflow production more than the forested and regenerating aspen uplands. Other studies within this region have focused on groundwater transmission between landscape units rather than on the whole catchment; Smerdon et al. (2005) studied upland-lake and lake-peatland linkages, Redding (2009) examined upland-peatland linkages and Ferone and Devito (2004) examined upland-peatland-pond connectivity. Further, Smerdon et al. (2005) simulated frequent seasonal reversals in groundwater flow direction between upland and lakes, which was controlled by the lakes' position in the catchment, high permeable sediments and lakebed deposits. Redding (2009) proposed a conceptual model outlining the direction of groundwater flow at an upland-peatland outwash site. He showed that the flow direction between forested upland and peatland was variable owing to different storage properties of peat and sandy outwash. Groundwater exchange between forested uplands and pond-peatland complexes in two fine-textured substrate basins varied seasonally, shifting direction between wet and dry seasons (Ferone and Devito, 2004). The flow reversals were controlled by available peatland storage and local flow system, as well as landscape position differences between the two study sites within moraine and lowland clay plain.

In summary, there has been much recent literature describing the hydrologic processes that control groundwater-aquifer interactions of the individual landscape units typical of the Boreal Plain. Groundwater transmission between landscape units, focused primarily on a few sites in north-central Alberta conducted at stand-level has also been studied. There has been relatively little focus on how these hydrologic processes interact to govern groundwater flow to streams in the Boreal Plain from a meso-scale perspective. Further work to understand groundwater-stream connectivity is needed to improve storage and flow routing parameterization for prediction of groundwater contributions to streamflow in the Boreal Plain. There is a pressing need to understand the role various landscape units play in regulating groundwater-stream interactions as this region is rapidly undergoing substantial land use change from energy and forestry sector disturbances (Price et al., 2010; Devito et al., 2011; Seitz et al., 2011) and climate change (Schindler and Donahue, 2006; Flanagan et al., 2011; Bergengren et al., 2011).

1.3 Research Objectives

The goal of this thesis is to improve the understanding of the roles large, valley-bottom peatlands, pine uplands and lakes have in regulating groundwater flows to Boreal Plain streams. The specific objectives of this study are to:

1) characterize the hydrogeology of a typical coarse-textured catchment;

2) determine groundwater recharge (to the water table) and groundwater exchange with the underlying sand aquifer for the landscape units that make up the catchment; and

3) determine how the landscape units interact to regulate groundwater contributions to the stream.

CHAPTER 2 - METHODS

2.1 Study Site Description

This research was conducted in the White Gull Creek Research Basin (WGCRB), which is a 603 km^2 catchment in the Boreal Plain ecozone of central Saskatchewan, Canada, ~50 km north of the town of Smeaton (Fig. 2.1). WGCRB falls within the BERMS (Boreal Ecosystem Research and Modeling Sites) research area, operated and funded by Environment Canada, Natural Resources Canada and Parks Canada from 1996 until 2012. The site is currently run by the University of Saskatchewan's Global Institute for Water Security.

The 1971-2000 climate of WGCRB was characterized by the Meteorological Service of Canada, via data collected at the Waskesiu Lake climate station, located 90 km west. Cold winters and moderately warm summers characterize the continental climate, and average monthly temperatures range from -17.9 to 16.2°C. Annual precipitation for the same period averages 467 mm, with 30% falling as snow. The catchment experienced cumulative dry moisture conditions from 2001-2003 countered by a large moisture surplus in 2004-2006 that was followed by the accumulation of several years of above mean precipitation (2006-2011; Fig. 2.2). The influence of total inter-annual precipitation and antecedent moisture conditions of the landscape is reflected in the different streamflow records across the WGCRB sub-catchments (Fig. 2.1). Little spatial variation in streamflow from the tributaries of White Gull Creek (WGC) occurs during wetter conditions, but flows in dry years are only persistent in the catchments underlain by glacial outwash. The study year, 2011, was considerably wetter than most years, as indicated by the three orders of magnitude increase in average streamflow recorded at the White Gull Creek hydrometric gauge (Water Survey of Canada (WSC) station 05KE010; Fig. 2.3).

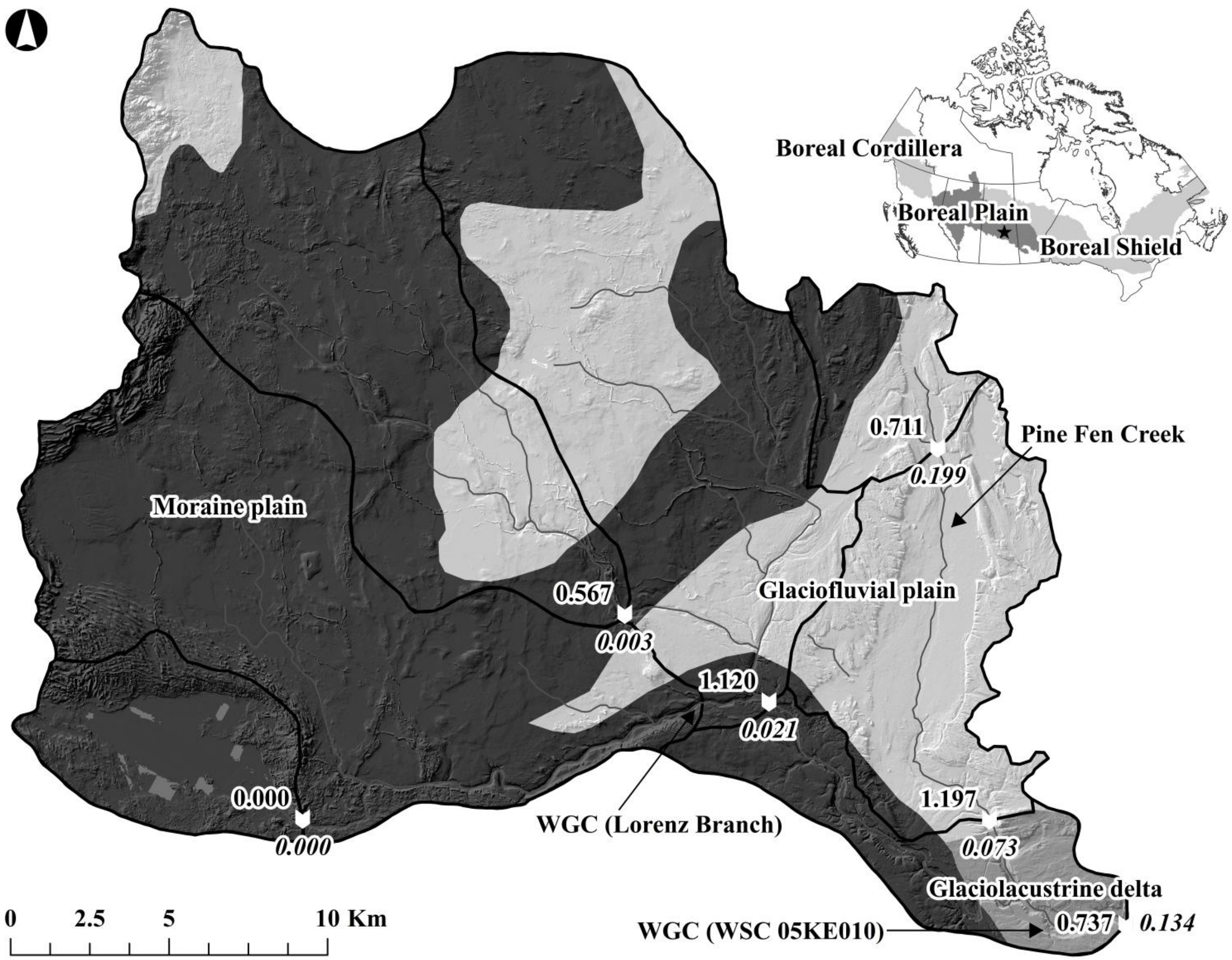


Figure 2.1: Surficial geology of White Gull Creek Research Basin (WGCRB) sub-catchments as delineated by Judd-Henrey et al. (2008) including: moraine plain (dark grey), glaciolacustrine delta (medium grey), and glaciofluvial plain (light grey). Values indicate measured streamflow (mm d^{-1}) during a dry (2003,italicized) and mesic year (2005, not italicized). Locations of streamflow measurement are indicated by white symbols. The inset panel depicts the extent of the Boreal forest of Canada showing the location of WGCRB (black star) within the Boreal Plain ecozone consisting of deep glacial deposits relative to the Boreal Shield ecozone to the east underlain by Canadian Shield bedrock and the Boreal Cordillera ecozone to the west located within the Canadian Rocky Mountains.

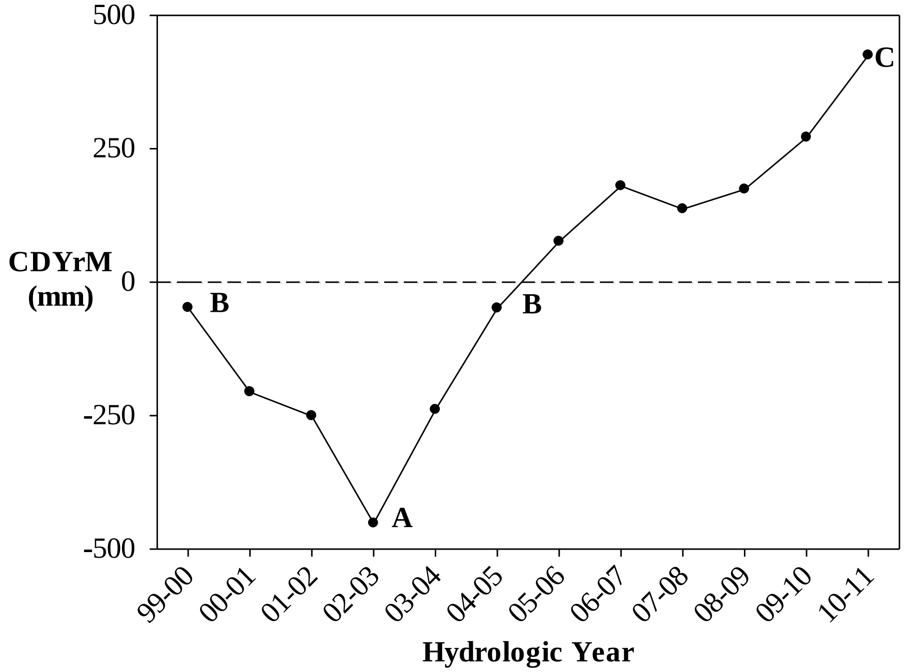


Figure 2.2: Cumulative departure of annual precipitation from the long-term yearly mean (CDYrM; 467 mm, 1971-2000 Waskesiu Lake climate station) over 12 hydrologic years measured at the mature jack pine stand (OJP) study site within WGCRB. Negative values indicate a cumulative moisture deficit and represents 'dry' or low antecedent moisture conditions (A), 'mesic' conditions (B) have net moisture deficit near zero, while accumulation above zero indicates potential moisture surplus and represents 'wet' or high antecedent moisture conditions (C).

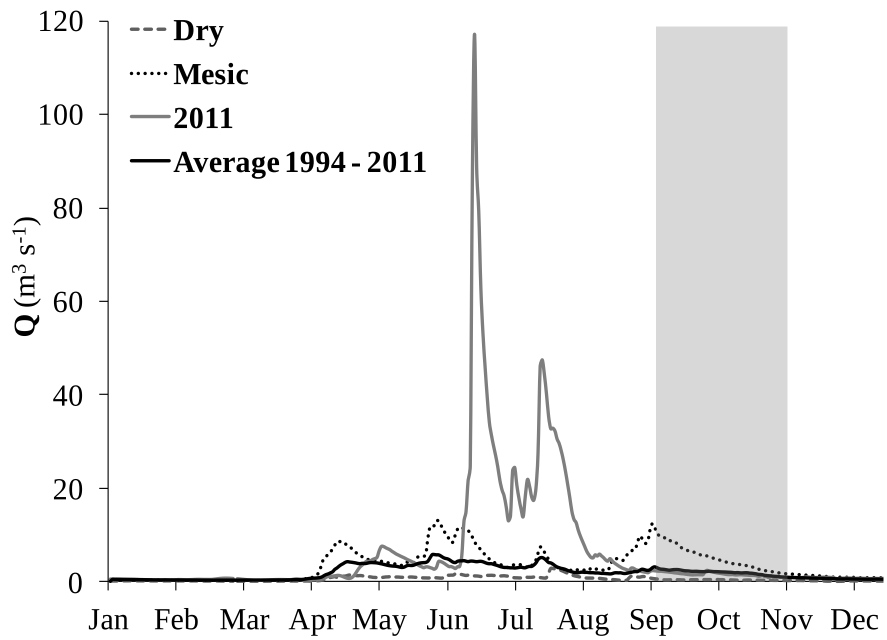


Figure 2.3: Mean daily hydrographs at White Gull Creek gauge WSC 05KE010 for a dry period (2001-2003; mean of 0.65 $m^3\ s^{-1}$), a mesic period (2004-2006; mean of 3.63 $m^3\ s^{-1}$), the period of record (1994 - 2011; mean of 1.96 $m^3\ s^{-1}$), and the wet study year (2011; mean of 6.19 $m^3\ s^{-1}$). The catchment has high inter-annual variability in streamflow. During the study period (indicated by the grey bar), 2011 streamflow was similar to the long term average flow.

Pine Fen Creek catchment (PFCC), a 96.7 km^2 peatland-dominated sub-catchment of WGCRB (Fig. 2.1) was selected for the investigation as its surficial geology is primarily coarse-textured outwash. PFCC is a gently rolling plain with a maximum topographic relief of 75 m, and land cover is a mix of lakes, peatland, and pine uplands (Fig. 2.4 and 2.5). The two lakes, Ispuchaw and Zeden, are connected by a small, peat-filled (~0.5 m deep) channel that has dimensions of ~125 m width x 200 m length. Ispuchaw Lake has a mean depth ~3 m with a maximum depth of ~6 m and transmits water to Zeden Lake, which has a mean depth ~6 m with a maximum depth of ~11 m. From manual probing, the lakes are likely gyttja-bottomed. There is a large (30.4 km^2), treed peatland in the valley bottom (classified as fen and unofficially named Pine Fen). Hydraulic and storage properties of Pine Fen located in the extreme north of the study site (classified as bog and fen) have been documented by Price and Fitzgibbon (1987). There is a second peatland in the northwest portion of the catchment, which straddles the divide (2.7 km^2 of which is in PFCC). Pine Fen Creek bisects Pine Fen from the north to south (Fig. 2.4 and 2.5). The creek continuously flows, even during dry years (Barr et al., 2012). The channel bed is composed of medium sand to cobbles, has a gradient of 0.002 m m^{-1}, and is frequently dammed by beaver at its headwaters and outlet (Fig. 2.4). There are two small streams that flow into the peatland. A groundwater spring discharges to the southwest side of the peatland (channel width of ~0.30 m) and Ispuchaw Lake discharges into the northeastern area of the peatland via a 1 m wide cobble-bed channel; only this second channel is a tributary to Pine Fen Creek.

The surficial geology of PFCC has been shaped by glacial processes. A predominant linear valley exists to the immediate northeast of the catchment and has been described as an incipient tunnel channel (Sjorgren et al., 2002; Judd-Henry et al., 2008). Although various hypothesis exist on the formation of tunnel channels (e.g. eroded due to pressurized sub-glacial waters), researchers agree that the termination of tunnel channels are often marked by large glaciofluvial fans containing outwash sediments (Cutler et al., 2002; van der Vegt et al., 2012). Boulder lags in the tunnel channel may have been transported onto the fan during an outburst flood of glacial meltwater, which could explain their presence near the surface of the upland and underlying the peatland. The fan was likely deposited on glaciolaucustrine substrate, which exists immediately south of the catchment (see Fig. 2.1). Price and Fitzgibbon (1987) characterized the surficial geology underlying the extreme northern extent of the peatland as a shallow (~1.2 - 2 m) sandy eolian and glaciofluvial layer underlain by compact clay-rich glacial till with low hydraulic conductivity (geometric mean 4 x 10^{-8} m s^{-1}).

Forest cover in PFCC is predominantly black spruce (*Picea mariana*) in lower relief depressions and the peatland. Lesser amounts of tamarack (*Larix laricina*) occur in wetter areas. Feather moss (*Pleurozium schreberi*) and mosses (*Sphagnum* spp.) with some bog cranberry (*Vaccinium vitis-idaea*) dominate ground cover in areas with increasing closed-canopies and/or greater soil moisture. Along the edges of open water, willow (*Salix* spp.), bog birch (*Betula glandifera*), and sedges (*Carex* spp.) are common. Paper birch (*Betula papyrifera*) and trembling aspen (*Populus tremuloides*) can be found in mesic areas, and mature or regenerating (i.e. clear-cut harvesting) jack pine (*Pinus banksiana*) occur in drier upland areas. Ground cover in open upland areas is dominated by *Cladina* spp., particularly reindeer lichen (*C. rangiferna*). As well, an overstory of Labrador tea (*Ledum groenlandicum*) occurs where trees do not form a closed-crown canopy. A tornado sheared upland trees in the extreme southwest of the catchment on 1 Aug 2011.

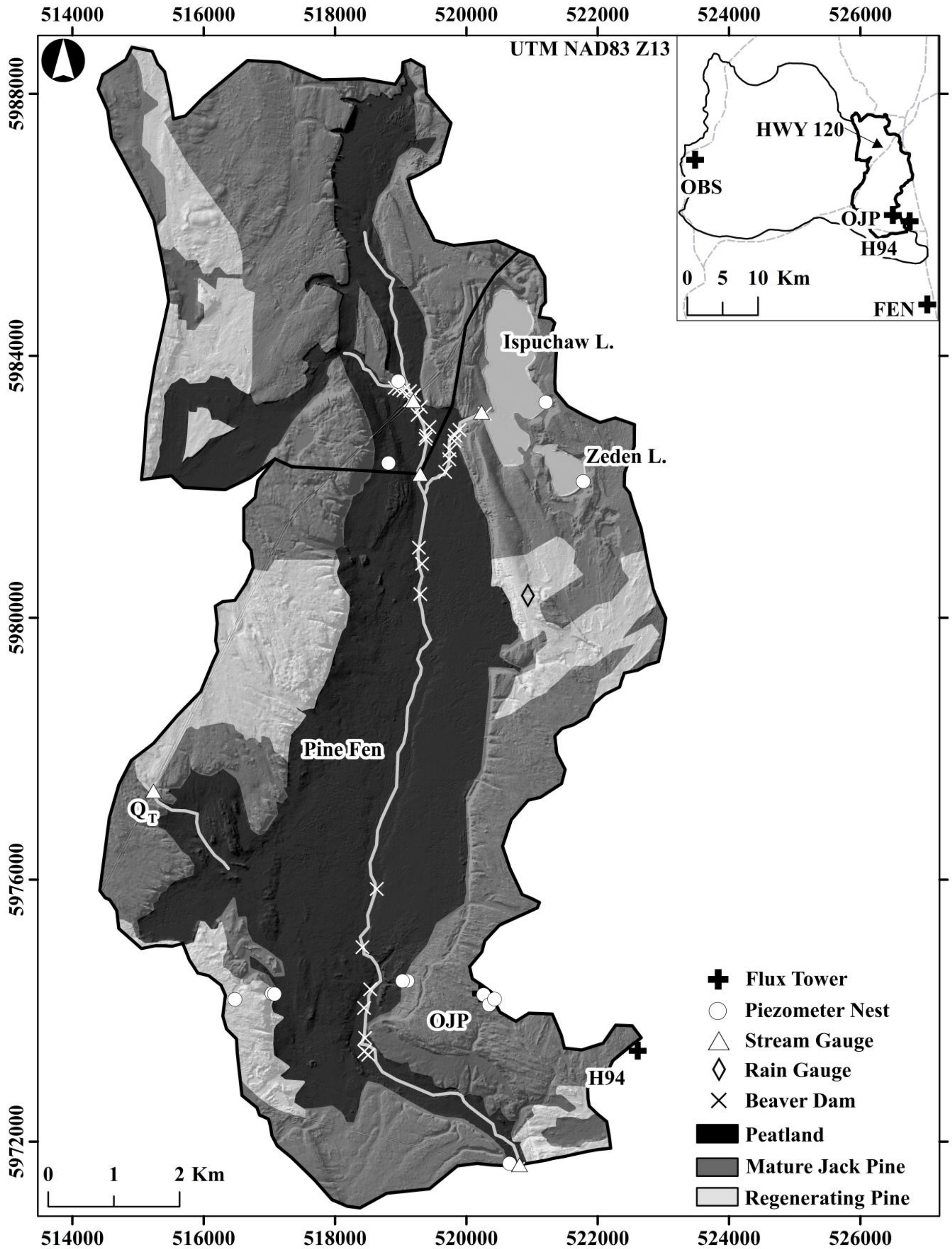


Figure 2.4: Pine Fen Creek sub-catchment (PFCC) delineated into four landscape unit types: lakes, peatland, mature and regenerating pine upland. Inset panel shows location of PFCC within WGCRB and four eddy covariance flux towers installed in a mature black spruce stand (OBS), mature jack pine stand (OJP), regenerating jack pine stand harvested in 1994 (H94), and *Carex* spp. dominated fen (Sandhill Fen; FEN). Q_T represents a tributary contributing surface and near-surface runoff to the peatland. The solid black line separates the upper (28.6 km^2) and lower (68.1 km^2) surface contributing areas of PFCC. Surficial geology is primarily coarse-textured outwash (see Fig. 2.1).

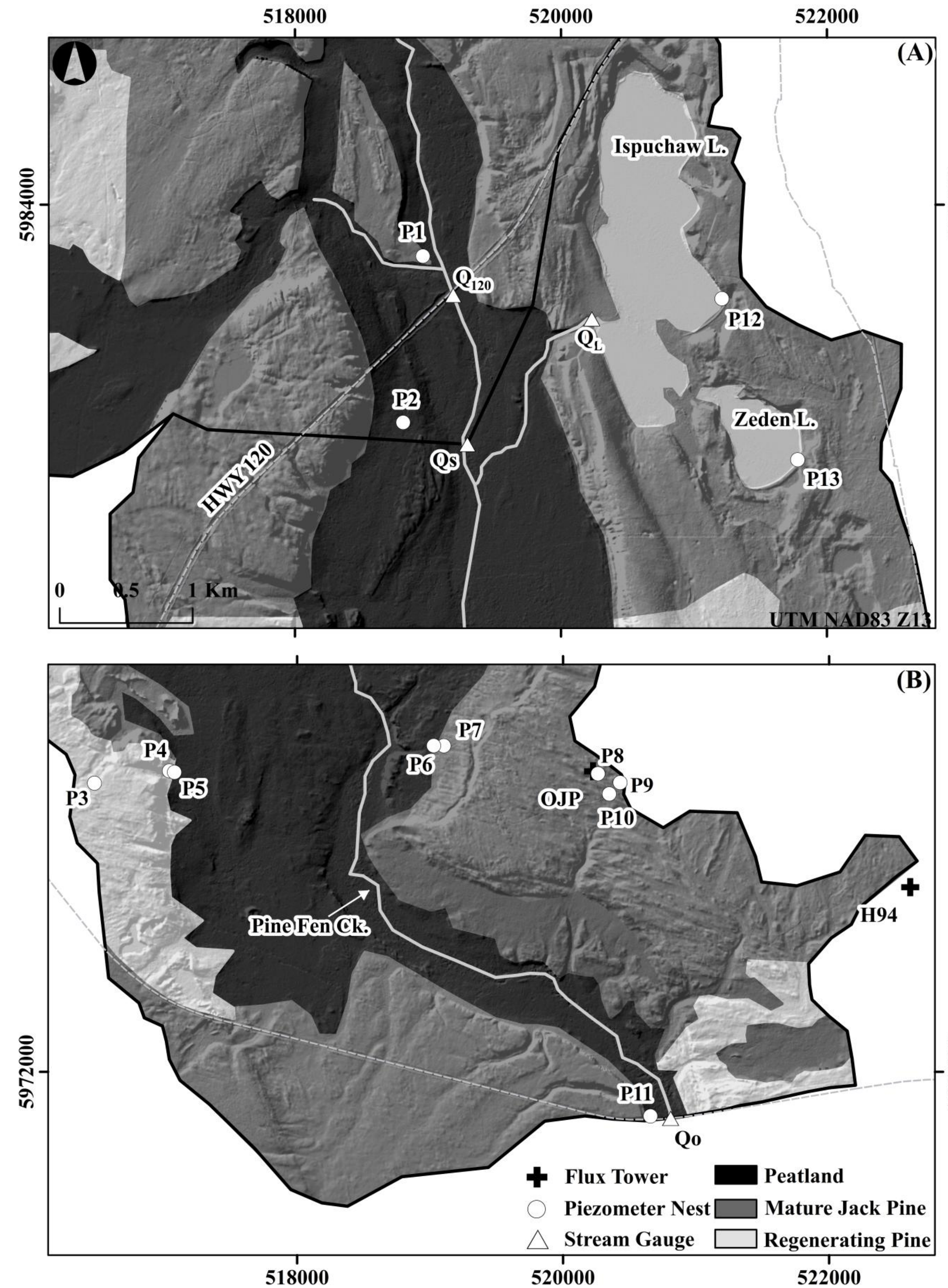


Figure 2.5: Location and ID of instrumentation in the northern (A) and southern (B) portions of PFCC.

Differences in relief, drainage, soil water holding capacity, and disturbance history have given rise to the heterogeneous land cover (i.e. landscape units) within the catchment (Fig. 2.4). Landscape units were delineated semi-manually using a combination of datasets and approaches. First, a maximum likelihood supervised classification was conducted using SPOT5 MS satellite images collected on September 24, 2008 with 10 m spatial resolution. There were 366 ground control points used for the supervised classification. These were marked during field surveys with a handheld global positioning system (GPS; Garmin etrex Legend; accuracy ±10 m). The resulting classification was improved with the Weyerhauser Saskatchewan Ltd. forest management plan map; and a 5 m x 5 m Light Detection and Ranging (LiDAR) digital elevation model (DEM). Landscape units delineated within PFCC (% of total catchment area) were: mature jack pine upland (48.7%); regenerating jack pine upland that was clear-cut between 2000 and 2004 (15.3%); lakes (1.8%); and peatland (34.2%; Fig. 2.6). Black spruce treed areas occurring on well drained sandy upland (6.0%) were not separated from the mature jack pine upland landscape unit as their interaction with the groundwater system was likely to be similar (Metcalfe and Buttle, 1999).

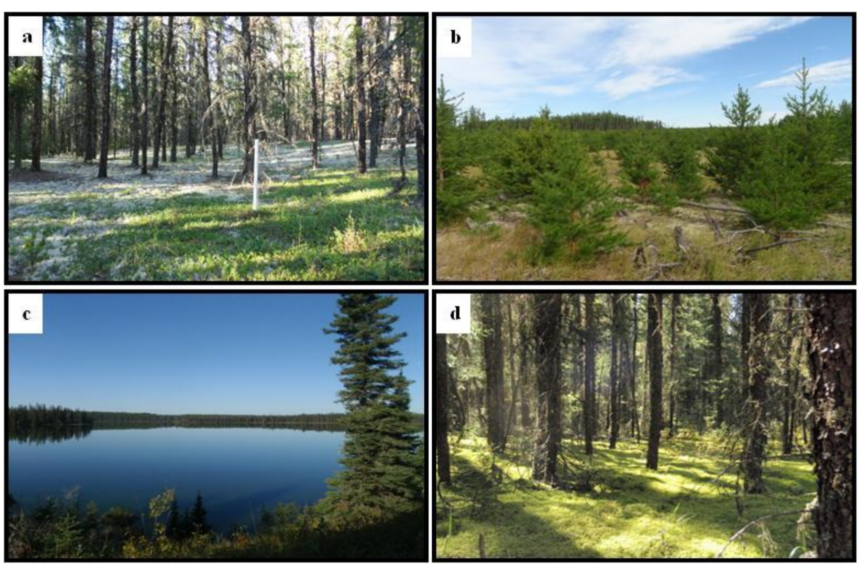


Figure 2.6: Photographs of the delineated landscape units: a) mature jack pine upland; b) regenerating pine upland; c) Ispuchaw Lake; and d) treed peatland.

2.2 Stratigraphic Characterization of Pine Fen Creek Catchment

To develop hydrogeological cross sections, the lithology of PFCC was investigated by drilling 11 boreholes with a Giddings drill rig. Borehole locations were chosen to complement existing borehole records (Judd-Henrey et al., 2008) within and near PFCC (see Appendix A). Boreholes ranged from 2 to 18 m. In addition, peat thickness was measured approximately every 40 m with an avalanche probe along four transects spanning the width of the peatland (n = 358). Bathymetry of Ispuchaw and Zeden Lakes was mapped by measuring water depth across two perpendicular transects in each lake using a weighted measuring tape. These data were corrected for the known lake level at the time of survey to give lake bed elevation.

2.3 Groundwater Instrumentation

Piezometer nests consisting of one shallow monitoring well and two piezometers were installed in each of the four landscape units to monitor groundwater flux (Fig. 2.4 and Appendix B), and complement existing monitoring locations. Shallow wells were constructed of fully slotted PVC pipe (0.025 m internal diameter, I.D.) and capped at the bottom. Piezometers were constructed of PVC (0.025 m I.D.) with 0.1 m slotted screens wrapped in drywall mesh and capped at the bottom. At four peatland sites, wells were pushed to the base of the organic column after pre-augering holes into the root mat using a smaller diameter auger than the PVC pipe. Adjacent to the wells, piezometers were installed 0.30-0.40 m above and 0.10-0.25 m below the peatland-sand interface. All PVC pipes were attached to a 2.1 m steel t-post with pipe clamps that were driven ~0.5 m into the underlying mineral soil with a hammer to achieve stability. The elevation of the peatland water table and peat surface were monitored relative to the t-post. At two mature pine upland sites and one regenerating pine upland site, piezometer nests were installed in sandy substrate that collapsed into the manually augured borehole below the water table. The PVC pipes were then driven through the loose sandy material to the bottom of the boreholes with a hammer. Shallow wells were installed to a depth of ~1.2 m below the surface. Piezometers consisted of slots at shallow (0.80-0.90 m) and deeper (1.1-1.4 m) depths. In each lake, piezometers were slotted at ~0.25 m and 0.70 m below the lakebed using similar installation techniques as in the pine upland. An additional deep monitoring well was installed in the regenerating pine upland to complete a transect across the catchment using three existing deep monitoring wells near the OJP flux tower site. All wells were constructed of PVC (0.052 I.D.) each having 1.5 m machine-slotted screens, capped at the bottom and installed into ~10 m boreholes pre-drilled using a Giddings rig.

All piezometer nests were developed using an inertial-lift tubing system (0.015 m Solinst foot valve connected to 2.5 m of tygon tubing) by removing at least three times the volume of water present after initial equilibration of the water level. Care was taken in the peatland to reduce 'over development' by not removing water until it was clear due to concerns of creating an artificial pore structure around the intake. The UTM coordinates of the piezometer network were surveyed with a handheld GPS (Garmin etrex Legend; accuracy ±10 m). The GPS coordinates of each nest were used to find ground elevation (m a.s.l.) from a 5 m x 5 m LiDAR DEM using ArcGIS as field surveying was not feasible.

Groundwater levels in each well and piezometer were measured daily from 7 September to 20 September and then every two weeks until 1 November using a piece of hollowed brass attached to a survey tape that would sound on contact with the water, hereafter referred to as a ‘plopper’ (G. van der Kamp, *personal communication*). These measurements were periodically corroborated with a small dial voltmeter connected to a length graded electric cable with two exposed wires that allowed an electric current to pass once they encountered water (Westbrook et al., 2006) and good agreement was found between the two methods. In addition, seven wells and piezometers were continuously measured half-hourly using submersible pressure transducers (Hobo U20 Water Leveloggers; accuracy ± 0.02 psi); data were corrected for barometric pressure (Solinst Barologger; accuracy ± 0.1 cm). Water levels were manually measured biweekly at each well throughout the study period to validate pressure transducer readings.

2.3.1 Delineating the PFCC Groundwater-shed

Contour plots of the water table were derived by the combination of kriging both measured and estimated point observations in Surfer version 10 (Golden Software Ltd.). This map was used to infer lateral groundwater flow patterns and delineate the PFCC groundwater-shed. Point observations included average measured water table elevation in wells over the study period, and average water levels in Pine Fen Creek, Ispuchaw Lake, Zeden Lake, and White Gull Creek. Additional water levels along Pine Fen Creek, White Gull Creek, and three ponds within PFCC were estimated with GPS coordinates and associated LiDAR elevation (created in 2006) to better constrain the water table. Estimated point observations reflect the moisture surplus conditions and relatively higher water levels of 2006 compared to previous dry years. Given that 2011 also had wet moisture conditions, using the LiDAR DEM to estimate water levels for point observations was reasonable. Krigged plots of the water table elevation showed that groundwater flow paths varied little throughout the short study period. The PFCC groundwater-shed was then delineated using the generalized groundwater flow paths, location of groundwater springs, lithology records, and topographic relief.

2.3.2 Groundwater Outflow from the Watershed

Lateral groundwater flow out of the watershed was estimated using:

$$G_{outlet} = -K_h \frac{dh}{dl} wb \quad (1)$$

where K_h is the geometric mean saturated hydraulic conductivity of the mineral and peat substrate (m s^{-1}; see Appendix B), d*h* is the change in hydraulic head measured across the surface watershed boundary (m) at the surface outflow gauging station and d*l* is the horizontal distance between the two measurement points (m), *w* is the width of the watershed at the point of outflow (m), *b* is the sum of average peat and sand aquifer thickness across the outlet (m).

2.4 Groundwater Recharge from the Pine Uplands and Peatland

Groundwater recharge is defined by Freeze and Cherry (1979) as "entry into the saturated zone of water made available at the water table surface, together with the associated flow away from the water table within the saturated zone". Both recharge across the water table (this section) and groundwater flow away or toward the water table between landscape units (hereafter exchange; section 2.5) were examined in this study.

Net groundwater recharge to the water table in the pine uplands and peatland was estimated at a point scale via the water table fluctuation (WTF) method (Healy and Cook, 2002; Cuthbert, 2010). This method is applicable to unconfined aquifers and assumes well locations to monitor water tables are representative of the landscape unit. The main limitations with the WTF method are difficulties in estimating specific yield for the study site and accurately accounting for the drainage term. A positive flux at the water table indicates conditions when the saturated groundwater zone is a sink for infiltrating water (i.e. groundwater recharge), and negative values indicate conditions when the saturated groundwater zone is a water source (herein referred to as drainage; Cuthbert et al., 2010). Accounting for drainage becomes more important where the depth to water table is large, creating a smoother hydrograph as a result of infiltrating rainfall moving through the unsaturated zone (Cuthbert, 2010). This method does not account for days when the rate of recharge equals the rate of drainage.

Net groundwater recharge in the mature (G_r^M) and regenerating (G_r^R) pine uplands were estimated in mm d^{-1} using (Cuthbert, 2010):

$$G_r^{LU} = \frac{S_y[(z_w(t) - z_w(t-1)]}{\Delta t} + \frac{K_h\,(\bar{H}^2 - b^2)}{L^2 - x^2} \tag{2}$$

where *LU* indicates the landscape unit type, $[z_w(t) - z_w(t\text{-}1)]$ is the daily change in unsaturated soil depth, z_w is the elevation of the water table measured in the crest upland well (m), and S_y is the specific yield, estimated at 0.30 at OJP by Barr et al. (2012). The drainage term includes K_h (m s^{-1}), which is the geometric mean hydraulic conductivity of the mineral substrate within the pine upland (see Appendix B). $\bar{H}$ is equal to the mean water table above the aquifer base during the study period (m), *b* is the thickness of the aquifer at the well (m), *L* is the distance between the well and a constant hydraulic head boundary taken to be Pine Fen Creek stage (m), and *x* is the distance between the well and groundwater divide (m). Net groundwater recharge was also calculated for the mature and regenerating pine upland toe slope wells. The shallow toe slope wells were located within the rooting zone of the jack pine (~2.5 m). G_r^M was obtained using the average daily groundwater recharge rate from the mature pine upland crest and toe slope wells. This method was also applied to calculate G_r^R using average rates from the regenerating pine upland.

Net groundwater recharge in the peatland was estimated by solving for G_r^P using Eq.2. A S_y of 0.30 was used, based on Price and Fitzgibbon's (1987) work on the upper 0.15 - 0.20 m in the extreme northern section of the peatland. G_r^P was averaged (area-weighted) across the four peatland wells.

2.5 Groundwater Exchange between Landscape Units and Sand Aquifer

Vertical and lateral hydraulic gradients were calculated from the observed hydraulic heads, and used to infer temporal variations in groundwater exchange with the underlying sand aquifer. Due to the inherent problems in scaling up from the point to landscape scale, net groundwater exchange for each of the four landscape units were also estimated using the water balance approach. The following sections describe these two approaches.

2.5.1 Hydraulic Gradients

Vertical hydraulic gradients (VHGs) were determined between piezometers via:

$$\boldsymbol{VHGs = -\,\mathrm{d}h/\mathrm{d}l} \tag{3}$$

where dh is the change in hydraulic head measured between two piezometers (m) and dl is the distance between the mid-point locations of the piezometer slots (m). VHGs were estimated between piezometers completed: 1) in the toe slope locations of mature and regenerating pine uplands; 2) in the lake bed and the lake level; and, 3) in the peat and underlying mineral sediment.

Lateral hydraulic gradients (LHGs) were determined between wells via:

$$\boldsymbol{LHGs = -\,\mathrm{d}h/\mathrm{d}l} \tag{4}$$

where dh is the change in hydraulic head measured between two wells (m) and dl is the horizontal distance between two wells (m). LHGs were estimated between: 1) upland toe slope and peatland wells; 2) lake levels and Pine Fen Creek stage; and, 3) peatland wells and Pine Fen Creek stage.

2.5.2 Groundwater Exchange via the Water Balance

Water balances for the mature and regenerating pine uplands, Ispuchaw and Zeden Lakes, and the peatland were estimated for the study period 7 September to 1 November 2011 using:

$$\boldsymbol{\Delta S = P - E + (Q_i - Q_o) + G^{LU}_{calc}} \tag{5}$$

where ΔS is change in storage, P is precipitation, E is open water evaporation (or evapotranspiration, ET, where appropriate). Runoff flowing into (Q_i) and out (Q_o) of the landscape unit refers to lateral surface flow (Devito et al., 2012). Eq. 5 was solved for net groundwater exchange, G_{calc}^{LU} (where *LU* indicates the landscape unit type), which refers to the net lateral and vertical flow within the subsurface. G_{calc}^{LU} includes the propagation of errors from each measured water balance term due to lack of an independent estimate of errors.

The G_{calc}^{LU} term refers to the total groundwater exchange of a given landscape unit with the underlying sand aquifer, other landscape units, and the stream in the lateral and vertical directions (Fig. 2.7). Interflow (i.e. unsaturated flow) between landscape units is included in the G_{calc}^{LU} term due to lack of measurement in this study. When calculating G_{calc}^{LU} for the peatland, groundwater exchange with the lakes was excluded due to the lack of physical connection between the two landscape units and the short duration of the study. Instead, the lakes were assumed to only exchange groundwater with the pine uplands and underlying sand aquifer (Fig. 2.7). Calculation of G_{calc}^{LU} for the pine uplands included groundwater inflow from beyond the surface watershed boundary, but within the groundwater-shed.

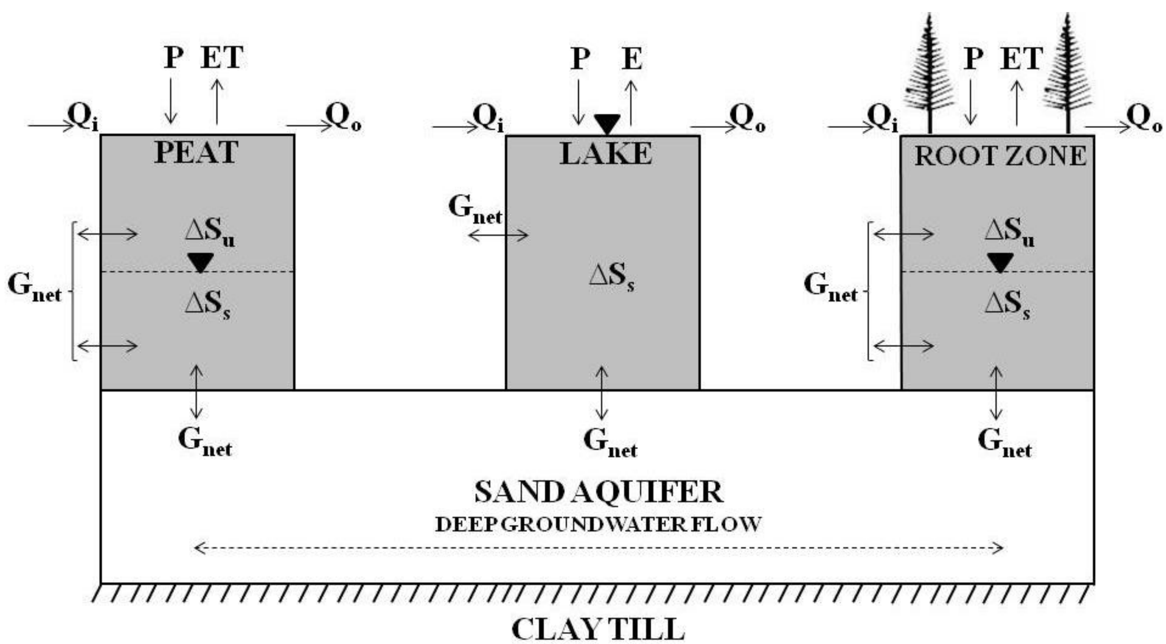


Figure 2.7: Conceptual diagram of control volumes (grey) to calculate landscape unit water balances. Arrows indicate direction of flux. The water table can be located within or below the root zone (of the pine upland). Water balance components are described in section 2.5.2. G_{net} is the sum of vertical and lateral groundwater exchange with the landscape unit and is represented in Eq. 5 as G_{calc}^{LU}.

The following sections (2.5.2.1 to 2.5.2.6) describe methods used to estimate each variable in Eq. 5.

2.5.2.1 Precipitation

Rainfall (P) was continuously measured with a recording tipping bucket rain gauge (Texas Electronics TR-525M) at 50 cm above ground in an open clearing. Accumulations were confirmed with a Meteorological Service of Canada Type B storage rain gauge at one site located in the sub-catchment (Fig. 2.4). The error observed between the cumulative precipitation measured by the tipping bucket and Type B rain gauge was 7%. Continual measurement of rainfall was also measured by Environment Canada at OJP using an accumulating gauge (Geonor with Alter shield) situated in the center of a small clearing. Daily precipitation was then estimated as an area weighted average of data from the two sites (mm; accuracy ± 15%, Winter et al., 1981). Forest interception was not taken into account due to the difficulty in accurately quantifying this flux over a large spatial area with different vegetation composition and structure.

2.5.2.2 Surface Inflow

There was no observed evidence of surface inflow to the lakes or peatland via the sandy uplands. This has been reported by others working in similar environments (e.g. Smerdon et al., 2005). Surface inflows to the peatland included the main Pine Fen Creek channel (Q_S), the western peatland tributary (Q_T), and Ispuchaw Lake outflow (Q_L) with areas of 25.7 km^2, 1.8 km^2, and 1.5 km^2, respectively (see Fig. 2.4). Total peatland inflow (Q_i) was the area-weighted sum of flow from Q_S, Q_T, and Q_L. Surface inflow to the peatland was periodically measured in the BERMS study ~1 km upstream of Q_S at Hwy 120 (Q_{120}; with an area of 27.3 km^2); however, several sizeable beaver dams were observed to regulate inflow at Hwy 120 during fall 2011. Therefore, Q_S was measured approximately 540 m downstream of the nearest beaver dam (Fig. 2.4). Both Q_S and Q_L were measured periodically using area-velocity methods with a Price Type

AA velocity meter as per Water Survey of Canada guidelines (Lane, 1999). Rating curves were developed for each site (see Appendix C) using manual discharge measurements and corresponding continuous measured half-hourly stream stage using pressure transducers, and corrected for barometric pressure as described in section 2.3. Transducers were installed in stilling wells that were secured to angle iron driven into the stream bed. Overbank flooding occurred at Q_S during the study period; therefore, velocity across the floodplain was taken to be equal to the channel velocity measurement at the bank. Discharge measurements were corroborated with six discrete measurements at Hwy 120. Q_{120} measurements were 50% to 82% lower than Q_S likely due to flow attenuation by beaver ponds (Westbrook et al., 2006). Q_T was measured daily using the volumetric method from 7 to 20 September and then every two weeks until 1 November. Estimated accuracy of Q_i is ± 30% (Winter et al., 1981).

2.5.2.3 Surface Outflow

Surface outflow was measured at the outlet of the peatland (Q_o), which was also the outlet of the catchment (see Fig. 2.4). Q_o has an area of 25.7 km^2. Manual discharge was periodically measured at the two culverts and stream stage was continuously measured half-hourly immediately upstream of the culverts using the same techniques described above. Beaver dams did not impede flow at the culverts during the study period. A rating curve was developed from observed stream discharge and continuous stream stage measurements (see Appendix C). Estimated accuracy of Q_o is ± 20% (Winter et al., 1981).

2.5.2.4 Evaporation

Open water evaporation (E) from Ispuchaw and Zeden Lakes was estimated using Granger and Hedstrom's (2011) newly developed hourly lake evaporation model (best estimate of accuracy ± 15%). The model relates wind speed over the lake surface and the horizontal gradient of land-lake temperature and vapour pressure contrast. Surface water temperature was measured hourly ~0.2 m below lake water level (Hobo temperature logger, Onset Corporation). Other model parameters not measured over the lake surface were measured at 2 m height at the FEN flux tower (~17 km south; see Fig. 2.4) including air temperature (HMP45 Vaisala probe, °C), relative humidity (HMP45 Vaisala probe, %), wind speed and direction (tri-axial sonic anemometer, Campbell Scientific, m s^{-1}). Fetch distances (m) were determined for each lake using a 45^{o} azimuth. Given that open water *E* is very sensitive to changes in wind speed, measured wind speed over the surface of Sandhill Fen (which was not treed) was used to estimate wind speed over the lake using methods described in Granger and Hedstrom (2010). Estimation of *E* using meteorological data measured at Sandhill Fen was reasonable as it was inundated with water and would experience less frictional resistance owing to the lack of trees.

2.5.2.5 Evapotranspiration

Evapotranspiration (*ET*) from the mature and regenerating pine upland were measured by Environment Canada (see Barr et al., 2012) with an eddy covariance (EC) system at OJP and H94, respectively (see Fig. 2.4). Flux measurements from OBS were used to estimate *ET* from the peatland, 26 km away, because of similar vegetation and depth to water table. The EC system consisted of a three dimensional sonic anemometer and a closed-path infrared gas analyzer (7000 LI-COR) located at twice the height of the forest canopy. Measurements of wind speed and latent water vapour content (Q_v, W m^{-2}) were taken at 10 Hz, calculated over a half hour period and corrected for the lack of surface energy balance closure (Barr et al., 2012). Actual *ET* was then calculated in mm d^{-1} via (accuracy ± 15%):

$$\boldsymbol{ET = \frac{Q_v}{\rho_w \cdot \lambda_v} \cdot k_c} \tag{6}$$

where ρ_w is the density of water (kg m^{-3}), λ_v is latent heat of vaporization (MJ kg^{-1}) and k_c is a unit converter from W m^{-2} d^{-1} to mm d^{-1} of 0.353.

2.5.2.6 Storage

The daily change in water storage (*ΔS*) in the mature and regenerating pine upland was estimated using the method outlined in Barr et al. (2012). *Δθ* was measured at OJP and H94 (see Fig. 2.4) using soil moisture reflectometers inserted vertically, at depths of 1.2 -1.5 m. Measures of water table position in the mature and regenerating pine uplands are described in section 2.3.

Change in storage in the peatland was calculated in mm d^{-1} using (accuracy ± 25 %, Spence and Woo 2006):

$$\Delta S = \Delta S_u + \Delta S_s = \Delta\theta\, [z(t) - z_w(t)] + S_y [z_w(t) - z_w(t-1)] \quad (7)$$

where ΔS_u is unsaturated storage (mm d^{-1}), ΔS_s is saturated storage change (mm d^{-1}), $\Delta\theta$ is change in volumetric soil moisture content (m m^{-3}), $z(t)$ is total soil depth (mm d^{-1}), and $z_w(t)$ is water table depth (mm d^{-1}). Storage was separated into portions of saturated and unsaturated because the water table in the peatland dropped below the topographic surface for the entire study period. For ΔS_u, $\Delta\theta$ was calculated from average daily soil moisture data recorded with calibrated soil moisture reflectometers (CS615, Campbell Scientific, mean of two profiles in units of m^3 m^{-3}) placed horizontally within peat, 7.5 cm below surface at OBS. A significant relationship (r^2 = 0.98, n = 32, p = <0.001) occurred between θ and the water table at OBS. This relationship was applied to the peatland as the water table did not drop more than 0.20 m below surface. Total saturated soil depth was the area-weighted sum of water table change measured at four wells completed in the peatland. Total soil depth was assumed to remain the same throughout the study period. Movement of the peat surface relative to a fixed datum (i.e. top of t-post) at each well location was small compared to the error in estimating peat surface elevation. S_y (0.30) was obtained from Price and Fitzgibbon (1987) in the upper 0.15 - 0.20 m of the extreme north of the peatland.

Open water storage (mm d^{-1}, accuracy ± 5%) was measured by continuously monitoring Ispuchaw and Zeden Lakes water level half hourly using barometrically corrected pressure transducers. Transducers were installed in stilling wells and water levels were observed at least every two weeks in each well to validate pressure transducer readings. Stilling wells were manually referenced to local benchmarks at the beginning and end of the study period using a survey level and rod to verify there was no movement.

2.5.3 Lake Winter Water Balance

Net groundwater exchange beneath Ispuchaw and Zeden Lakes was also estimated using the winter water balance method over winter 2010-2011 (Welsh et al., 2012). This technique was used to calculate an average net groundwater exchange rate for the whole lake as the uncertainty associated with using the few piezometers installed was high due to the inherent spatial variability in lakebed permeability, and the lack of a meteorological station on a lake. With the lakes ice covered, the only variables used to compute the net groundwater flux in the winter water balance were precipitation (as snow), and surface inflow or outflow. Lake levels were surveyed at ice-augered holes using a survey level and rod. A snow survey was completed from the lake centers to edge (n = 72) following the methods described in Pomeroy and Gray (1995). Mean snow water equivalent (*SWE*) was determined from snow depth measurements and snow density samples taken with an Eastern Snow Conference snow sampler. Ispuchaw Lake was not likely to have surface outflow during winter due to freezing at the channel outlet; however, surface outflow was not ground-truthed during the 2011 winter. Assuming no surface inflow or outflow and a constant lake surface area, the average winter net groundwater exchange rate (mm d^{-1}) for Ispuchaw Lake (G^{I}_{meas}) and Zeden Lake (G^{Z}_{meas}) was determined via Welsh et al. (2012):

$$G_{meas}^{LU} = [(z_2 - z_1) - SWE]/(t_2 - t_1) \quad (8)$$

where *LU* indicates the lake, z_2 is the lake level (m a.s.l.) immediately before the onset of lake ice (November 2010), z_1 is the lake level (m a.s.l.) before the snow load has melted or the break-up of lake ice (March 2011), *SWE* is the maximum snow load on the lake surface (mm), and (t_2 - t_1) is the time between lake level surveys (128 days). Lakes were considered net groundwater discharge sites if G_{meas}^{LU} > 0.05 mm d^{-1}, net groundwater recharge sites when G_{meas}^{LU} < -0.05 mm d^{-1}, or flow-through if the inflow was approximately equal to the outflow (G_{meas}^{LU} = 0 within ± 0.05 mm d^{-1}; Welsh et al., 2012).

2.6 Groundwater Contributions to Streamflow

2.6.1 Rainfall - Runoff Analysis

Stream hydrographs were separated using a recession curve (as outlined in Chapman, 1999 and Spence, 2006) to calculate event runoff (Q_e, m d^{-1}). Q_e was estimated as the sum of the daily difference between the calculated flow on day t (Q_t; assuming no event) and observed stream discharge (Q) until $Q_t > Q$:

$$Q_t = Q_o e^{-t/t^*} \quad (9)$$

where Q_o is the streamflow on the day before the event and t^* is the recession coefficient calculated using the reciprocal of the slope of the best fit regression line between ln(Q) and t for the falling limb of the hydrograph before the event. Event rainfall volume, P_e, was estimated as the product of total event P and the contributing area to compare volume runoff ratios (Q_e/P_e).

The hydrological function of the peatland at a specific time and place was determined using the method outlined in Spence (2007). Briefly, the peatland was considered to be predominately storing water if the runoff rate was lower than the change in storage $\Delta S/\Delta t$ (mm d^{-1}) whereas the opposite was true if the peatland was predominately discharging runoff. A discharging peatland was then considered to be either: 1) contributing water if the internally generated runoff (outflow - inflow) was greater than inflow; or 2) transmitting water if the internally generated runoff was equal to or less than inflow (sensu Spence et al., 2011).

2.6.2 Scaled Landscape Unit Outflows

To gain a broader understanding of how landscape units interact to regulate groundwater contributions to streams in other wet and dry years, landscape unit outflows in PFCC were estimated over 10 hydrologic years (2001-2011) using methods described in Barr et al. (2012). This method used flux tower measurements collected within different landscape units to calculate vertical water balances and estimate stand-level outflows scaled to the area-weighted land cover fraction within WGCRB. Outflows were defined, as per Barr et al. (2012), as the sum of (lateral) surface runoff, interflow and groundwater flow. For this thesis, many of the same data described in Barr et al. (2012) were used to construct the vertical water balances in PFCC, as it is a sub-catchment of WGCRB. However, other data more specific to the landscape units in PFCC, when available, were used, and these are described below.

Stand-level outflows were estimated annually (beginning October 1 and ending September 31 of the following year) to reduce uncertainties related to temporal lags in lateral groundwater transmission. Outflows were calculated for the mature and regenerating pine uplands, lakes and peatland via (Barr et al., 2012):

$$\boldsymbol{O = P - ET - \Delta S} \tag{10}$$

where P is the rate of precipitation at OJP (mm y^{-1}) located in PFCC, ET is the rate of evapotranspiration at OJP, H94 and OBS for the mature pine upland, regenerating pine upland and peatland, respectively (mm y^{-1}). Estimates of lake evaporation were based on average values from a Saskatchewan Boreal Plain lake presented in Granger and Hedstrom (2011). ΔS is the rate of storage change measured at OJP, H94 and OBS (mm y^{-1}). Lake storage was estimated using water level measurements at Ispuchaw Lake and was assumed to be the same for Zeden Lake.

The long term water table record measured in the mature pine upland crest well (2002-2011) was used to estimate annual groundwater recession rates via the WTF method (Healy and Cook, 2002) to corroborate pine upland outflows.

Stand-level outflows (mm y^{-1}) were scaled to the area-weighted land cover fraction (%) within the groundwater contributing area above the (surface) catchment outlet (8.54×10^7 m^2) to provide more realistic estimates of groundwater contributions to streamflow. Land cover fraction was 45%, 13%, 40%, and 2% for the mature and regenerating pine uplands, peatland and lakes, respectively.

At the catchment scale, four single-day field measurements of streamflow (mm d^{-1}) taken in October 2003, 2005, 2009 and 2011 were used to corroborate indirect mean daily estimates of streamflow (scaled catchment streamflow) computed via the sum of scaled landscape unit outflows.

2.6.3 Water Isotope Analysis

Water samples were analyzed for isotopic composition to corroborate hydrometric measures of groundwater contribution from each landscape unit to streamflow. Rain was sampled from a standard rain gauge located at Sandhill Fen during and immediately following the cessation of the 11 to 13 September rainfall event. Water was also collected daily from Q_{120} and Q_o for 7 to 20 September and then every two weeks until 1 November. Before and after the 11 to 13 September rainfall event, water was collected from deep pine upland groundwater wells ($G_{sand(7m)}$ at ~7 m below ground surface), shallow pine upland groundwater wells and piezometers completed in mineral soil underlying the peatland ($G_{sand(1m)}$ at ~1 m below ground surface), peatland wells (G_{peat}), White Gull Creek (Lorenz branch, see Fig. 2.1), Zeden Lake and Ispuchaw Lake. Groundwater was collected using an inertial-lift tubing system (Solinst, Ontario). Wells and piezometers were purged approximately three well volumes prior to sample collection. Stream samples and lake samples were collected at approximately one half depth of the water column. Lake samples were collected every two weeks after the large rainfall event approximately 1 m from shore or in the lake center. Water samples for isotope analysis were collected in 20 mL scintillation vials without air space, capped tightly to avoid evaporation, and stored at room temperature. Samples were analyzed for the stable isotopes of ^{18}O and ^{2}H at the National Hydrology Research Centre, Saskatoon, Canada using spectroscopy following the methods described by Lis et al. (2008). Isotopic composition is expressed in terms of $^{2}H/^{1}H$ and $^{18}O/^{16}O$ ratios and represented by delta notation (δ) in units per mill (‰) relative to Standard Mean Ocean Water (SMOW). Specific conductance, *EC* (μs cm^{-1}) was measured in the field on a separate sample using a hand held meter (sensION156, Hach Instruments).

CHAPTER 3 - RESULTS

3.1 Hydrogeological Setting

The glacially derived surficial sediments of PFCC overly a bedrock surface composed of marine shales and sandstones belonging to the Manville Group which lay ~100 m below the ground surface (Judd-Henrey et al., 2008). Five cross-sections through PFCC were produced from the hydrogeological investigation (Figs. 3.1 and 3.2) and show that the surficial geology of the catchment consists of an outwash plain, mainly composed of medium-coarse sand, which is underlain primarily by clay-rich glacial till with some silt, sand and pebbles. Borehole records indicate that sand thickness underlying the brunisolic soil (<0.1 m) varies from 2 to >18 m in the pine uplands (Figs. 3.2a and 3.2c). The depth to glacial till at the north end of the catchment (transect A-A'), decreases from west to east (Fig. 3.2a). The ridge located ~3500 m along transect B-B' (Fig. 3.2b) was previously classified as a moraine (Campbell and Simpson, 1986) and was thus not ground-truthed as part of this thesis. At transect C-C', sand was found to be >18 m thick at the east margin of the peatland (Fig. 3.2c; drilling limited by number of drill extensions). At transect D-D', the sand aquifer is thin, with only 3 m of sand underlying the peat (Fig. 3.2d). The longitudinal cross section that extends outside the catchment boundary indicates there is 70 m of sand in the north (Figs. 3.1 and 3.2e), decreasing to an average of 6 m near the peatland inlet at Hwy 120.

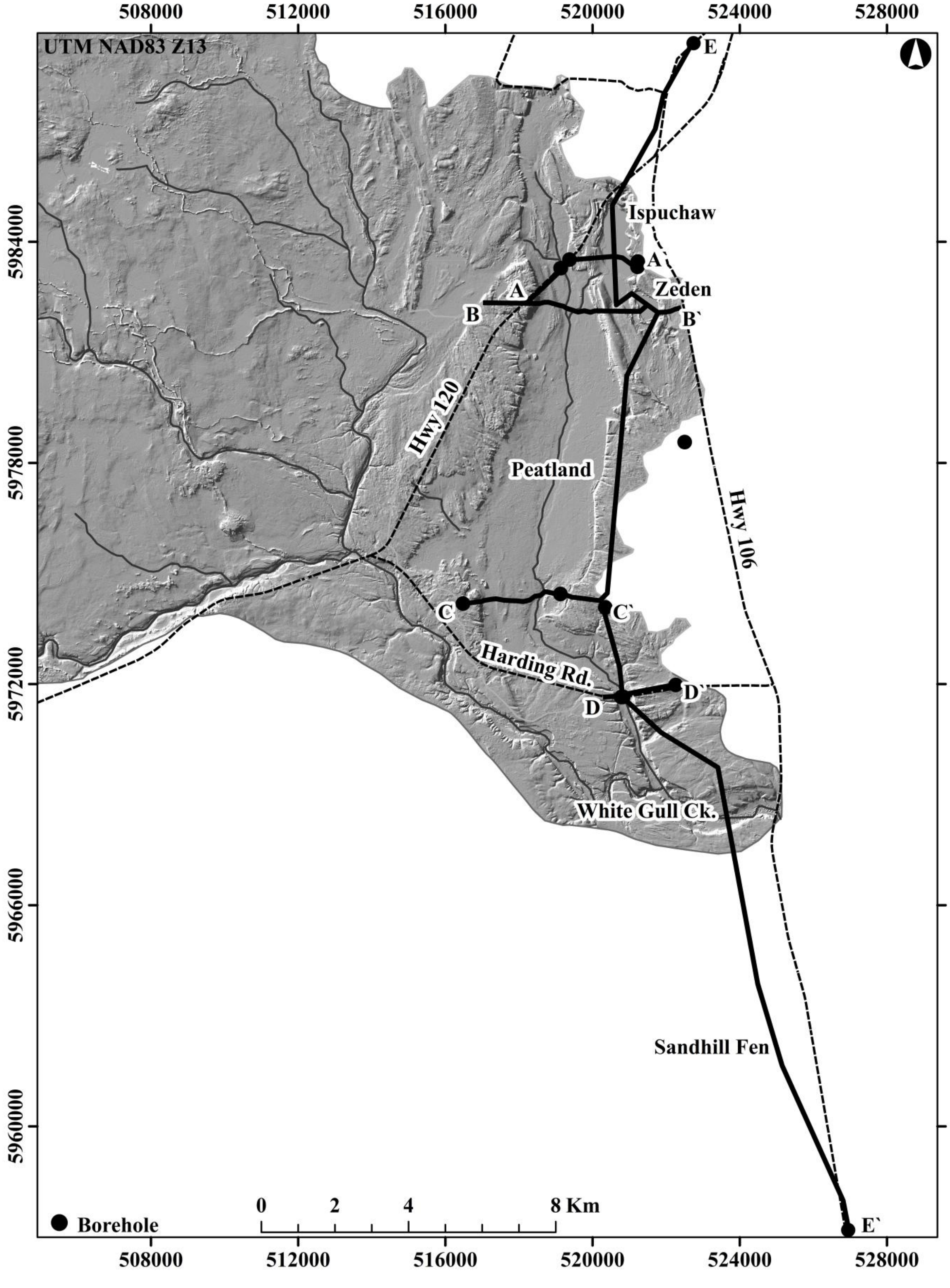
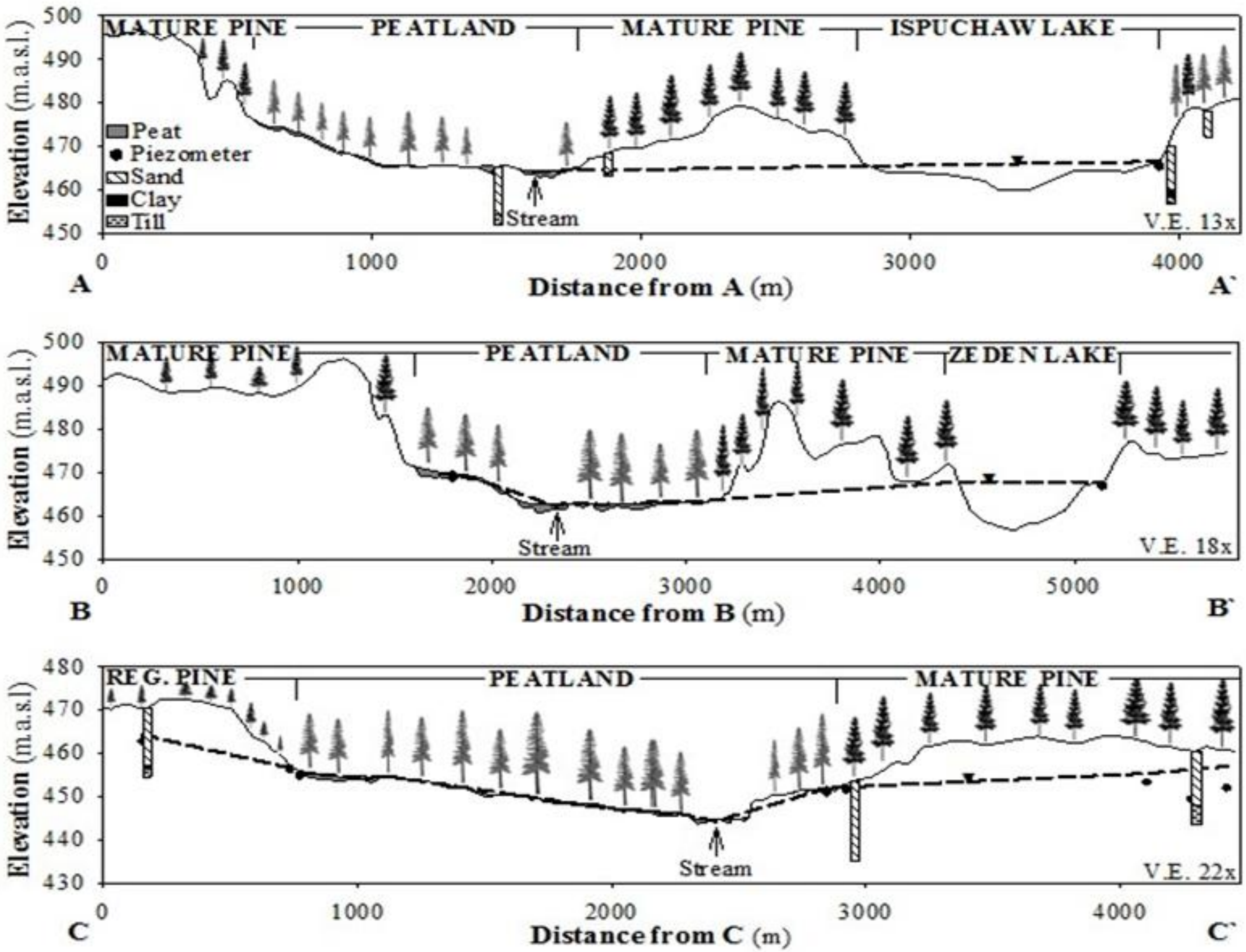


Figure 3.1: Map representing the locations of five hydrogeological cross-sections through PFCC including locations of boreholes.



MATURE PINE
PEATLAND
MATURE PINE
ISPUCHAW LAKE
Peat
Piezometer
Sand
Clay
Till
Stream
V.E. 13x
Elevation (m.a.s.l.)
Distance from A (m)
A
A`
MATURE PINE
PEATLAND
MATURE PINE
ZEDEN LAKE
Stream
V.E. 18x
Elevation (m.a.s.l.)
Distance from B (m)
B
B`
REG. PINE
PEATLAND
MATURE PINE
Stream
V.E. 22x
Elevation (m.a.s.l)
Distance from C (m)
C
C`

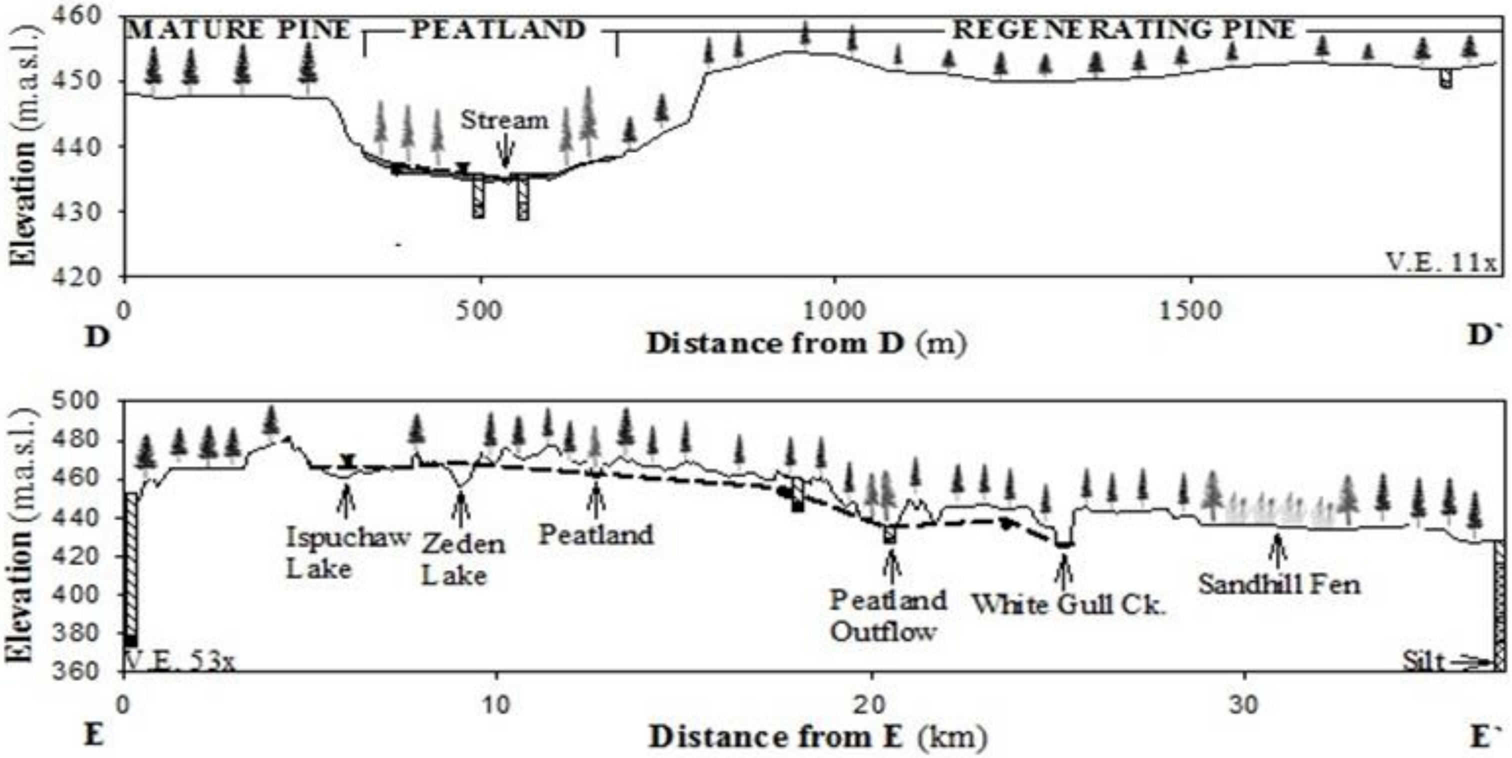


Figure 3.2: Hydrogeological cross-sections through PFCC: a) A-A', b) B-B', c) C-C', d) D-D', and e) E-E'. The dotted black line indicates water table elevation during the time of transect mapping. Water table elevation delineated between piezometers, lake levels, and stream stages were interpreted based on LHGs.

The peatland has an average thickness of 0.65 m, a maximum depth of ~2.0 m, and was ice-free between mid-August and late October in 2011. Peat thickness generally decreases toward the south end of the catchment. The peatland has a mean slope of 0.0025 m m^{-1}, and a typical microrelief of hummocks and hollows that have an average elevation difference in the order of ± 0.25 m. Peat stratigraphy is characterized by a surficial fibric layer, graded H2 – H4 on the von Post (1922) scale (Fig. 3.3a). Below this (to ~0.6-0.7 cm depth) the peat becomes mesic, and then gradually becomes more humified with greater depth. The stratum beneath the peat layer consists of well-sorted medium sand; however, sections of gravel, cobbles, and boulders were observed at the peat-sand interface (Fig. 3.3b).

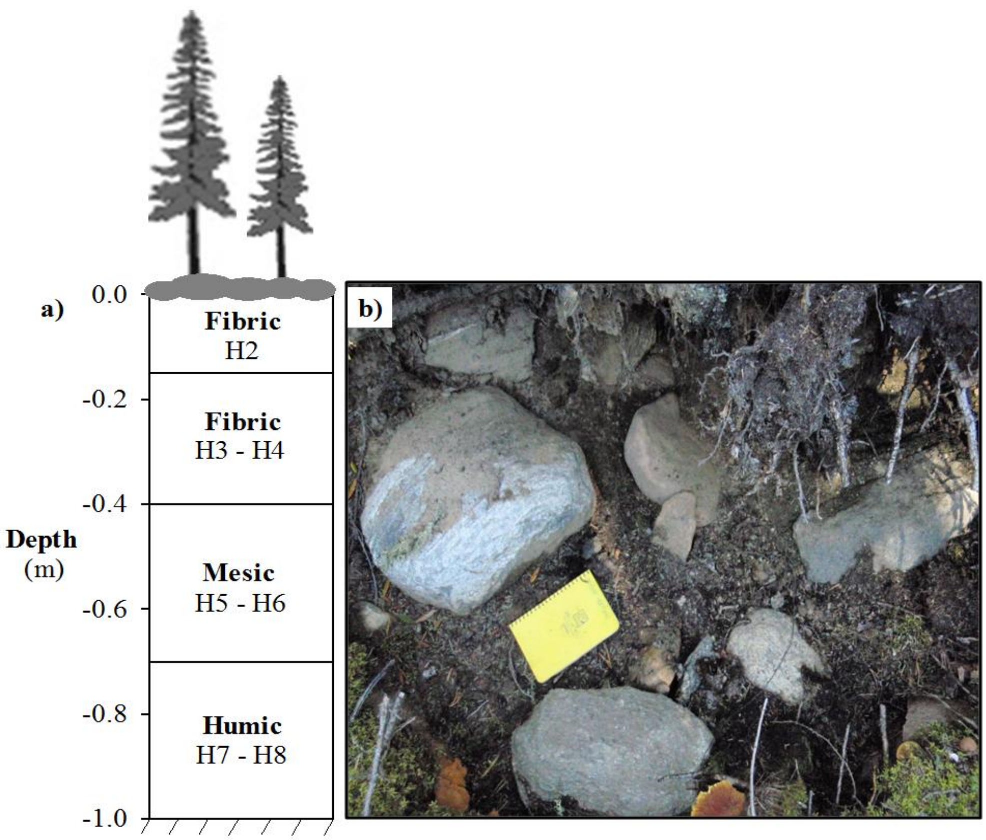


Figure 3.3: a) Cartoon representation of peat stratigraphy characterized during the installation of the piezometer network using the von Post (1922) scale of decomposition. Black spruce, sphagnum and feather mosses dominate the peatland cover; b) Photo of boulders and cobbles underlying peat at a location ~2300 m along transect C-C' (Figs. 3.1 and 3.2c).

Both Zeden (0.3 km^2) and Ispuchaw (1.5 km^2) Lakes have near shore lakebed sediments consisting of coarse sand. Some cobbles were observed at Ispuchaw Lake. A borehole drilled on the shore of Ispuchaw Lake revealed clay (glacial till) beneath 10 m of sand (Fig. 3.2a).

The delineated groundwater-shed of PFCC is 89.5 km^2 (Fig. 3.4), which is 7% smaller than the surface watershed. Along the west and east sides of the groundwater-shed, the groundwater divide lies within the extent of the surficial sand aquifer. The boundary generally follows the surface catchment estimated based on topography and the location of groundwater springs and seepages. The long term record of water tables in the triangularly arranged piezometers at OJP (Fig. 2.4) corroborates the direction of groundwater flow toward Pine Fen Creek throughout the study period. Equipotential lines indicate the groundwater divide does not extend as far north as the surface catchment boundary. The northern groundwater divide proved more difficult to delineate due to fewer measurements of hydraulic head. Estimation of the northern divide within the peatland does corroborate with the surficial geology terrain classification map and indicates that the divide generally falls on a glaciofluvial-moraine plain interface (Campbell and Simpson, 1986). The groundwater-shed also extends past the catchment outlet (4.1 km^2).

Equipotential lines show that the main vector of groundwater flow is from north to south, and that the stream gains groundwater along its length. Groundwater flow paths indicate down valley flow toward the stream from the pine uplands and lakes; however, the path of groundwater flow is intercepted by the peatland before it reaches the stream. Groundwater loss beneath the surface watershed boundary and streamflow gauging location at the catchment outlet was estimated to be 1 x 10^3 m^3 d^{-1} during the study period. The loss represents only 2% of the mean daily catchment surface outflow and was considered negligible.

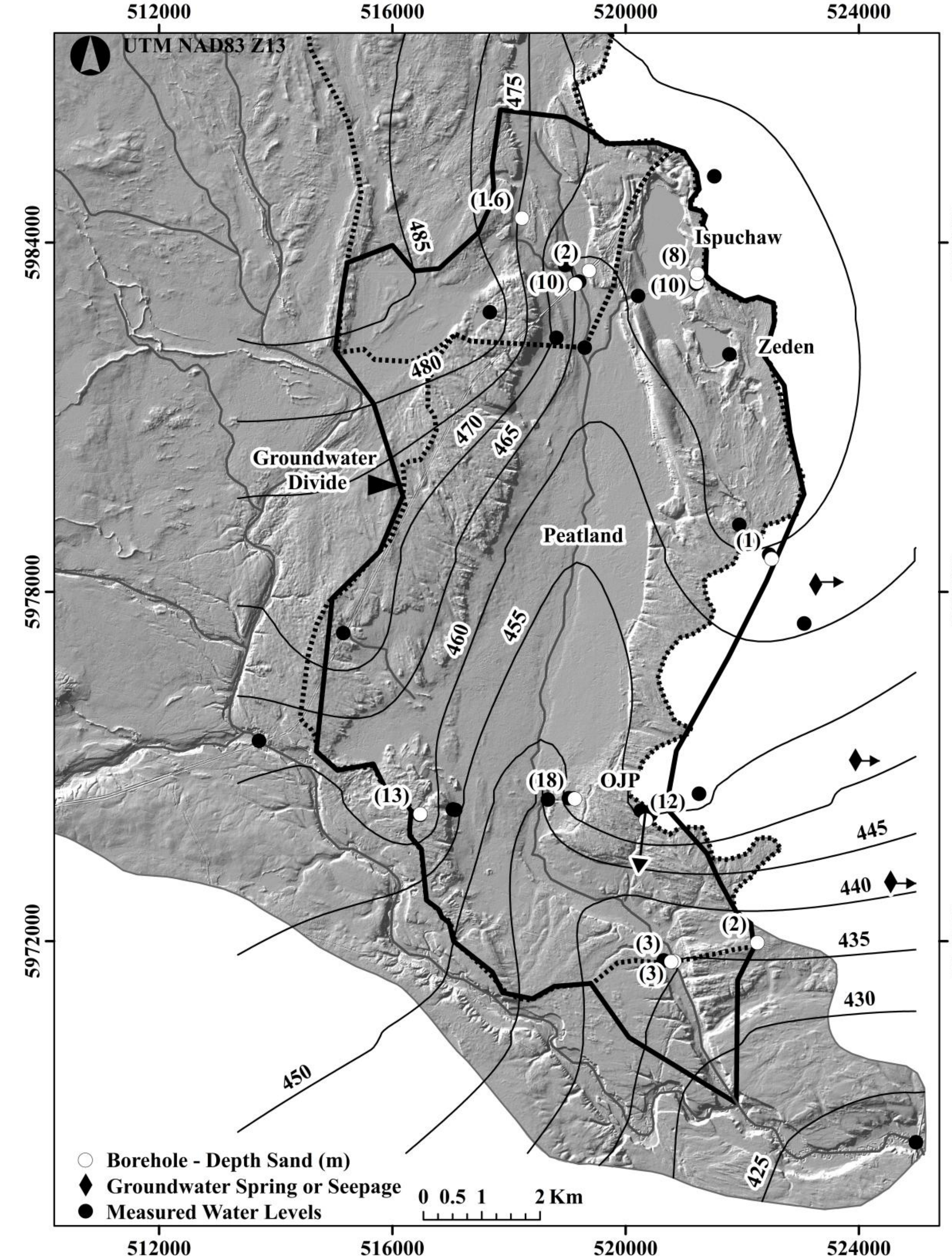


Figure 3.4: Delineated groundwater-shed of PFCC (solid black line) with the surface catchment boundary (dashed black line) overlain on a LiDAR DEM; locations of observed groundwater springs or seepage (arrow indicated flow direction), measured water table elevation (m) during the study period, and depth of sand from borehole records (in brackets) are also shown. Black arrow at OJP site indicates direction of groundwater flow using triangulation method.

3.2 Groundwater Recharge from the Pine Uplands and Peatland

There was greater drainage away from the mature pine upland than recharge to its water table over the 56 day study period (G_r^M = -8.8 mm). Average net groundwater drainage rates from the mature pine upland increased from -2.3 to -6.5 mm d^{-1} during September to October, respectively. In contrast, the regenerating pine upland experienced net groundwater recharge to its water table (G_r^R = 2.8 mm). However, average net groundwater recharge rates decreased from September (0.2 mm d^{-1}) to October (0.02 mm d^{-1}). Average net groundwater recharge was near zero in the peatland (0.1 mm d^{-1}) and ranged from -3.4 to 7.6 mm d^{-1} over the course of the study period.

3.3 Groundwater Exchange between Landscape Units and Sand Aquifer

3.3.1 Temporal Variation in Hydraulic Gradients

VHGs in the mature pine upland (toe slope) indicated groundwater flow away from the water table (0.01 to 0.04 m m^{-1}) toward the underlying sand aquifer (Fig. 3.5). In contrast, VHGs in the regenerating pine upland (toe slope) indicated upwards groundwater flow toward the water table (-0.08 to -0.01 m m^{-1}).

VHGs between the lake stage and lakebed indicated flow reversals at both Ispuchaw Lake (-0.11 to 0.04 m m^{-1}) and Zeden Lake (-0.10 to 0.03 m m^{-1}) on short time scales (Fig. 3.5). Both lakes primarily gained groundwater from the underlying sand aquifer following the 11 to 13 September rainfall event.

The peatland had variable groundwater exchange with the underlying sand aquifer (Fig. 3.5). The west peatland nest (P5) indicated flow was downward toward the underlying sand aquifer (0.02 to 0.08 m m^{-1}), whereas the south peatland nest (P11) showed groundwater flow was upward into the peat along a steeper gradient (-0.37 to -0.05 m m^{-1}). Flow reversals occurred at the east peatland nest (P6) throughout the study period (-0.11 to 0.09 m m^{-1}).

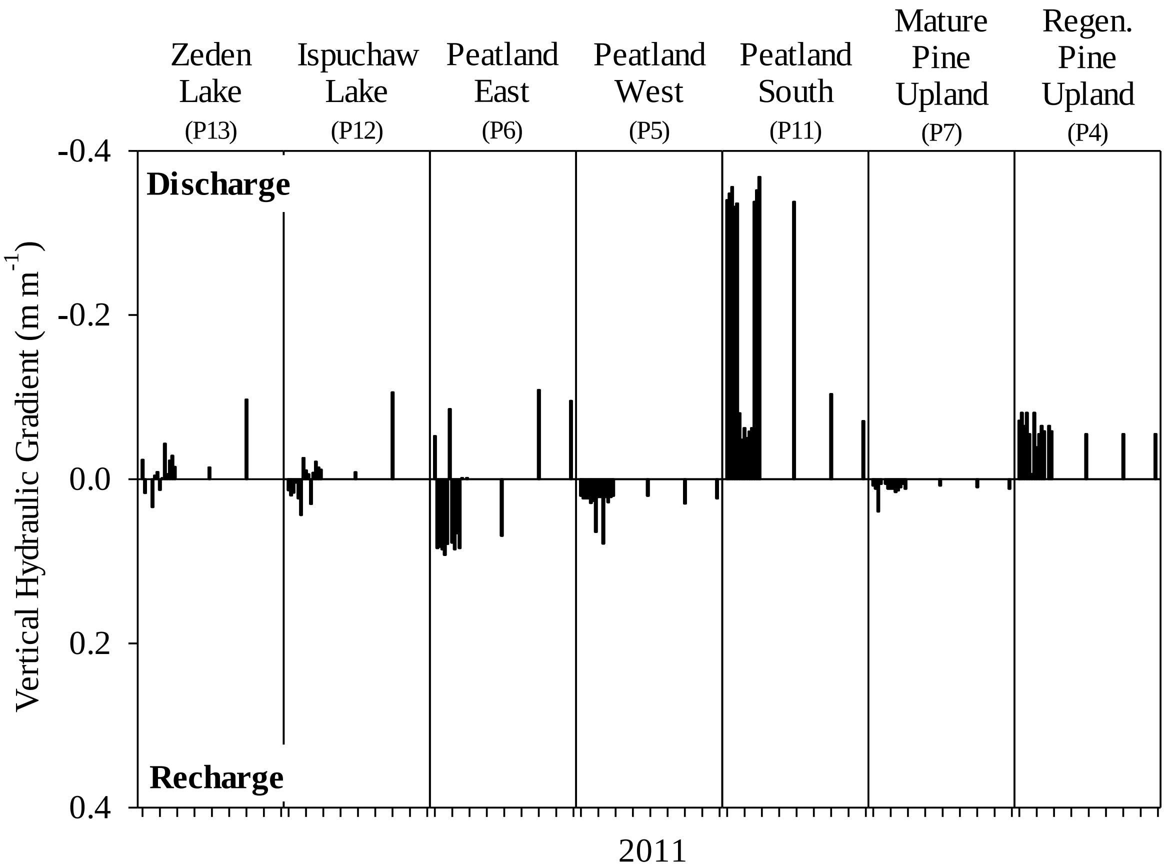


Figure 3.5: Vertical hydraulic head gradients (dimensionless) for piezometers completed in different landscape units: lakes (below the base of the lake), peatland (peatland-underlying sand aquifer interface), and upland (within ~1.3 m depth from surface). Bars are daily values of VHGs for 7 to 20 September, 4 October, 19 October, and 1 November 2011. Discharge conditions indicate upward groundwater flow into the landscape unit and plot above the zero gradient line; recharge conditions indicate groundwater flow into the underlying sand aquifer and plot below.

Groundwater flow direction was consistently from the pine uplands toward the peatland over the course of the study period. LHGs remained stable and were steeper in the regenerating pine upland (0.01 - 0.02 m m^{-1}) compared to the mature pine upland (0.005 m m^{-1}).

Consistent and gentle LHGs (5.0 x 10^{-4} m m^{-1}) indicate that groundwater flow was from Ispuchaw Lake to Zeden Lake. LHGs from the lakes to Pine Fen Creek were stable and similar (0.005 m m^{-1}) over the course of the study period.

LHGs across the peatland indicate groundwater converged at Pine Fen Creek throughout the study period. LHGs from peatland wells to Pine Fen Creek remained stable and were the same from the north (P2) and south (P11) peatland wells to the stream (0.01 m m^{-1}), but less steep from west (P5; 0.005 m m^{-1}) and east (P6; 0.005 m m^{-1}) peatland wells to the stream.

3.3.2 Groundwater Exchange with the Sand Aquifer

Spatial variation in net groundwater exchange within PFCC was the result of variation in the water balance terms other than *P* during the 7 September to 1 November 2011 study period (Table 3.1).

Table 3.1: Cumulative water balance components for landscape units within PFCC over the 56 day study period. Values are presented in m^3 (mm) per landscape unit area[1]. Note that values are rounded to 2 significant figures.

Landscape Unit		Area (m^2)	P [2]	E	Q_i	Q_o	ΔS	G_{calc}^{LU} [3]
Pine Uplands	Mature	3.04×10^7	1.5×10^6 *(48)*	1.7×10^6 *(55)*	0	0	1.9×10^4 *(0.6)*	2.3×10^5 *(7.4)*
	Regen.	1.02×10^7	4.9×10^5 *(48)*	5.0×10^5 *(49)*	0	0	1.1×10^5 *(11)*	1.0×10^5 *(12)*
Lakes	Ispuchaw	1.47×10^6	7.0×10^4 *(48)*	1.7×10^5 *(120)*	0	1.3×10^5 *(85)*	-1.3×10^4 *(-8.6)*	2.2×10^5 *(150)*
	Zeden	2.89×10^5	1.4×10^4 *(48)*	3.9×10^4 *(140)*	0	0	-8.5×10^3 *(-29)*	1.7×10^4 *(59)*
Peatland		2.57×10^7	1.2×10^6 *(48)*	1.6×10^6 *(61)*	1.7×10^6 *(67)*	2.0×10^6 *(79)*	3.3×10^4 *(1.3)*	6.7×10^5 *(26)*

[1]Landscape unit area was defined as the area within the lower contributing area of the surface watershed (6.81×10^7 m^2).

[2] Values were not corrected for forest interception.

[3] G_{calc}^{LU} was estimated using Eq. 5. A positive value indicates the landscape unit had a net gain of groundwater flow from the underlying sand aquifer.

Rainfall during the study period was typical for fall, based on the 1971-2000 climate normal from Waskesiu Lake station. However, rainfall recorded in June-July 2011 was almost 70% greater than the climate normal.

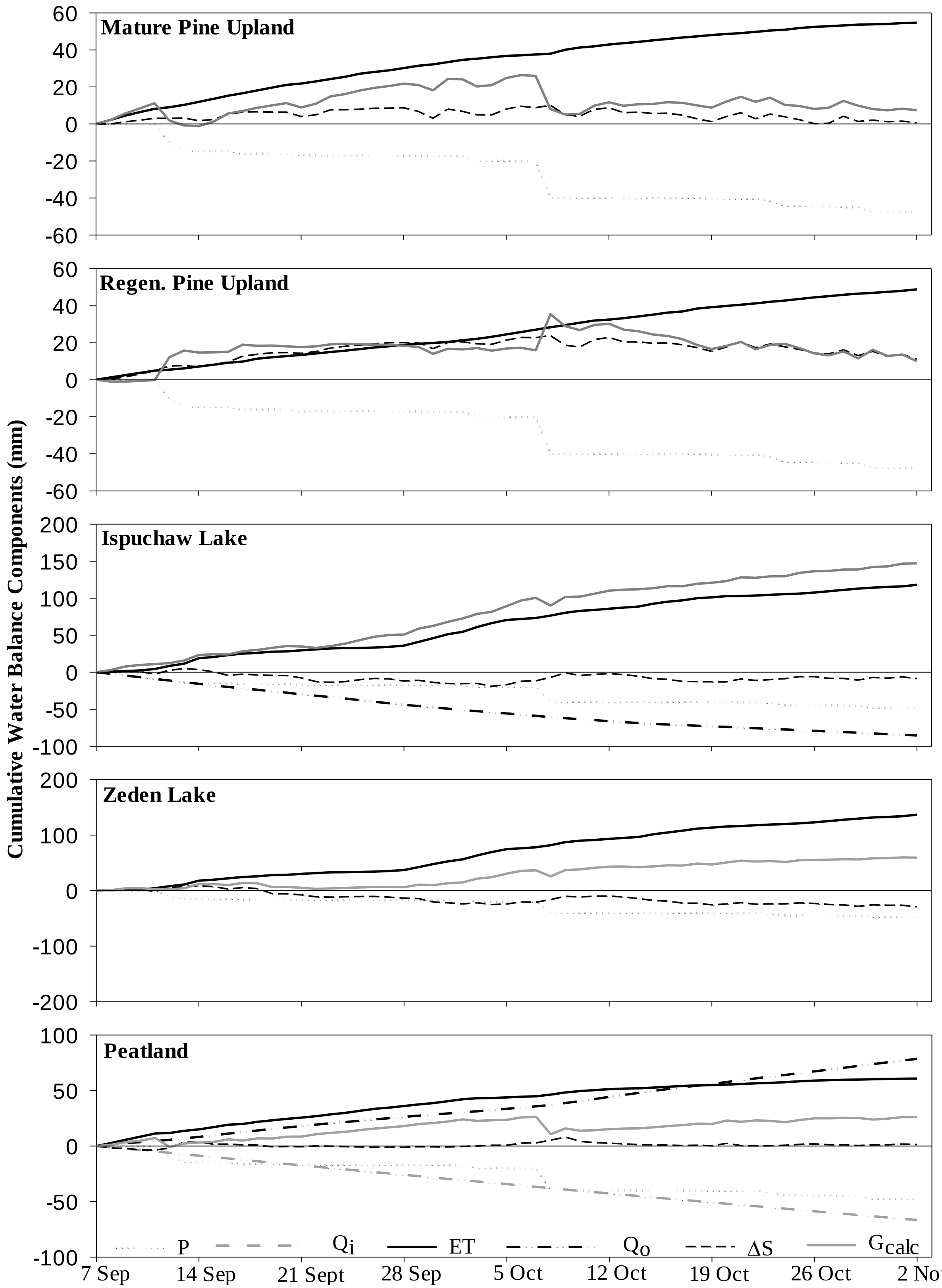


Figure 3.6: Comparison of cumulative water balances for each landscape unit in mm per unit area (lower surface contributing area) for 2011 study period. Values are relative to land surface (0).

No surface inflow to the lakes or peatland via the sandy uplands was observed during rainfall events. Total peatland inflow (Q_i) had a mean daily rate of 1.2 mm d^{-1}. Q_T contributed only 0.1% of average Q_i over the study period. Daily Q_L steadily decreased from 11% to 5% of Q_i by the end of the study period. Ispuchaw lake level was nearly constant throughout the study period, declining only 9 mm. Outflowing water flowed through an old beaver dam at the outlet. Q_o was the largest peatland water flux over the study period (Table 3.1), with a mean daily rate of 1.4 mm d^{-1}.

E was a dominant hydrologic flux for both Ispuchaw and Zeden Lakes during the study period (Table 3.1). The overall mean daily *E* rate was 2.1 mm d^{-1} for Ispuchaw Lake and 2.4 mm d^{-1} for Zeden Lake. Daily mean *ET* in the mature and regenerating pine uplands was 1.0 and 0.9 mm d^{-1}, respectively. Resulting *ET* over the study period was greater in the mature pine upland compared to the regenerating pine upland (Table 3.1). The peatland showed a greater daily mean *ET* rate of 1.1 mm d^{-1}. The first seasonal frost occurred between the 13-15 September when air temperatures dropped to 0°C. Coincident with this, the transition of conifers into winter dormancy was evident from the observed attenuation of the diurnal groundwater level fluctuations.

Outputs exceeded inputs at both the pine uplands and lakes resulting in storage deficits in both landscape units by the end of the study period (Table 3.1). The regenerating pine upland had a greater overall storage loss compared to the mature pine upland. Zeden Lake had a greater total storage loss than Ispuchaw Lake. In contrast, the peatland had a cumulative storage gain over the study period (Fig. 3.6).

All landscape units had an overall net gain of groundwater from the underlying sand aquifer over the 56 day study period (Table 3.1; Fig. 3.6). Although the mature pine uplands showed a net gain, mean monthly G_{calc}^{M} ranged from 1.0 mm d^{-1} to -0.5 mm d^{-1} in September and October, respectively. A similar trend was observed in the regenerating pine upland where G_{calc}^{R} varied from 0.7 mm d^{-1} (September) to -0.1 mm d^{-1} (October). Average net groundwater exchange rates indicate Ispuchaw and Zeden Lakes experienced a net gain of groundwater over the fall study period (Table 3.2), but over-winter groundwater exchange rates indicate Ispuchaw (-1.4 mm d^{-1}) and Zeden (-1.2 mm d^{-1}) Lakes lost water to the underlying aquifer. The peatland had an overall net gain of groundwater flow with estimated G_{calc}^{P} ranging from -15 to 5.1 mm d^{-1}. Events of large net loss of groundwater from the peatland corresponded with large rainfall events (e.g. -15 mm d^{-1} on 7 October; Fig. 3.9).

Table 3.2: Summary of average net groundwater exchange for each landscape unit during the 7 Sep to 1 Nov 2011 study period; a positive value indicates the landscape unit had a net gain of groundwater flow from the underlying sand aquifer.

Landscape Unit		**Net Groundwater Exchange** Volume (m^3 d^{-1})	**Net Groundwater Exchange** Rate (mm d^{-1})
Pine Uplands	Mature	4.0×10^3	0.1
	Regen.	1.8×10^3	0.2
Lakes	Ispuchaw	3.9×10^3	2.6
	Zeden	3.0×10^2	1.1
Peatland		1.2×10^4	0.5

3.4 Groundwater Contributions to Streamflow

3.4.1 Rainfall-Runoff Analysis

Shallow groundwater responded within an hour of the two large rainfall events that occurred on 11 to 13 September and 7 October (Table 3.3). The upland toe slope water table responded faster to rainfall inputs (quicker time to peak) compared to the peatland (Table 3.3). Continuous water table records indicate the slope of hydrograph recession was consistently steeper in an upland toe slope well (P7) compared to a peatland well (P6; Fig. 3.7). The peatland water table receded 5 to 10% slower than the hillslope after rainfall events. Water tables in wells not continuously measured show similar patterns.

Water table response occurred at the time of effective water input for all peatland wells; however, lag time to peak water table elevation varied between wells. For both rainfall events, water tables recorded in the peatland wells rose to within ~15 cm from the peat surface (Table 3.3). This water level pattern was not recorded in the northern peatland well (P2) located near a ridge of peat (see Fig. 2.5).

Table 3.3: Response analysis for two events during fall 2011 study period. P6, P11, and P2 indicate the east, south, and north peatland wells, respectively. P7 indicates the eastern upland toe slope well (see Fig. 2.5).

Date	Storm Event		Groundwater Response										
	P (mm)	P_i (mm hr^{-1})	Lag (hr)			Total Rise (cm)				Depth Blw. Grd. Surface (cm)			
			P6	P11	P2	P6	P11	P2	P7	P6	P11	P2	P7
12-Sep	15	0.4	6	3.5	5.5	3	4	4	11	13	15	39	55
7-Oct	20	1.5	6	2.5	7.5	5	5	11	24	11	10	31	42

[1]rainfall intensity (P_i) was determined by dividing total rainfall by duration of event: 40 hrs (11 to 13 September) and 13 hrs (7 October).
[2]centroid lag-to-peak is the time between the weighted mean time of rainfall input and the peak
[3]total rise indicates the difference in elevation from the beginning of hydrograph rise to peak
[4]surface elevations are 452.05 (P6), 437.53 (P11), 469.72 (P2) and 452.90 (P7) m a.s.l.
[5]Distances between peatland wells and adjacent toe slopes are 30 m (P6), 50 m (P11) and 250 m (P2).

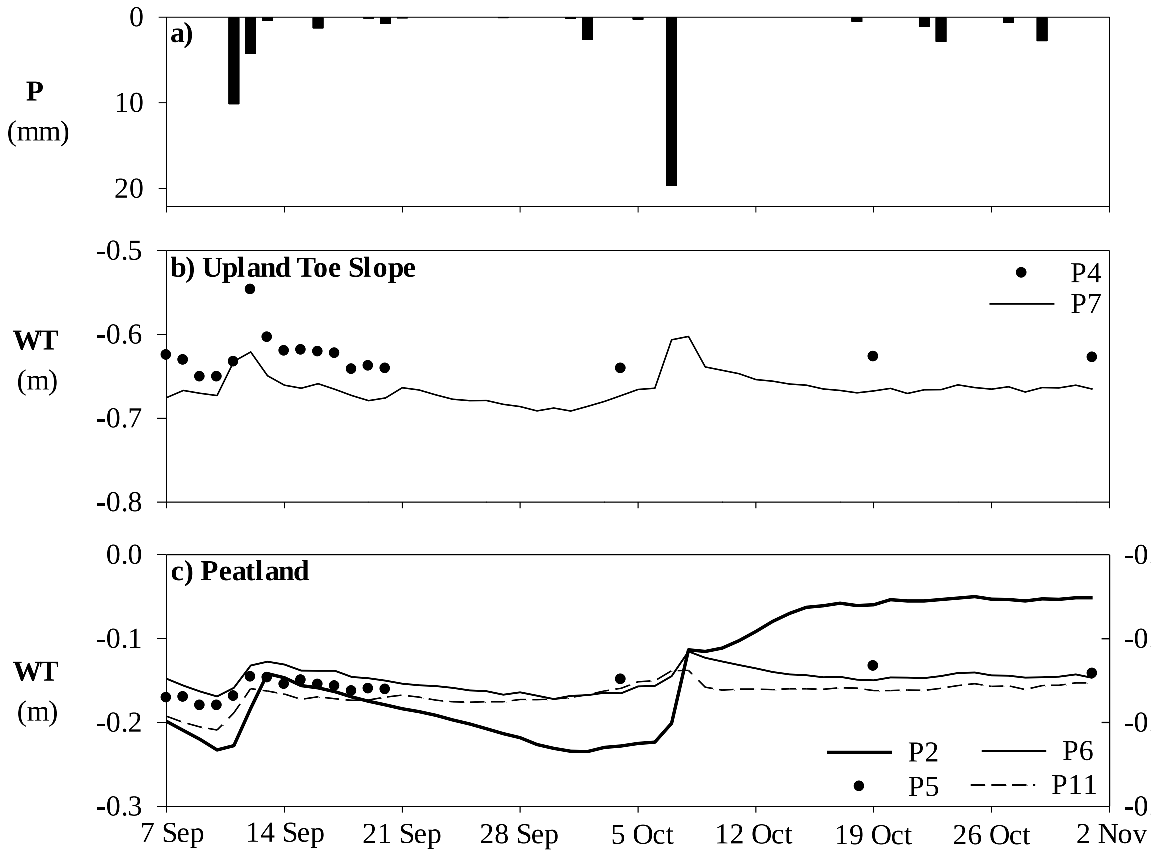


Figure 3.7: a) Rainfall recorded over the study period in 2011; b) Water table elevation recorded in the east (P5) and west (P4) upland toe slope wells; c) Water table elevation recorded in the east (P6), west (P5), south (P11), and north (P2) peatland wells (see Fig. 2.5). The water table record for P11 is associated with the secondary axis, whereas all other well records are indicated on the primary axis.

The estimated time for a peak flood wave to travel from the peatland inlet to outlet was ~141 hours using a mean stream cross-section area of 13 m^2, a mean stream velocity of 0.025 m s^{-1}, and the stream length of 13.9 km. The centroid-lag to peak discharge determined at the peatland outlet was less than the time estimated for the peak flood wave to arrive (Table 3.4). Peak discharge was larger for Q_o compared to Q_i for both events (Fig. 3.8b). Q_i was relatively stable, ranging from 0.330 to 0.380 m^3 s^{-1}, whereas Q_o was more responsive to events, varying from 0.292 to 0.552 m^3 s^{-1} (Fig. 3.8b). The 15 mm rainfall event on 11 to 13 September released ~1900 m^3 and ~1.67 x 10^5 m^3 of stored water at the inlet and outlet, respectively. The volume runoff ratio was 0.003 at the inlet (t^* = 204 hr) and 0.16 at the outlet (t^* = 13.7 hr).

Table 3.4: Response analysis for two events during the 2011 study period for the peatland surface inflow (Q_i) and outflow (Q_o).

Date	Storm Event		Stream Response			
	P (mm)	P_i (mm hr^{-1})	Lag (hr)		Peak Q (m^3 s^{-1})	
			Q_i	Q_o	Q_i	Q_o
12-Sep	15	0.4	1	51	0.389	0.433
07-Oct	20	1.5	22	82	0.365	0.556

For 8 to 11 September, Q_o showed a steady recession to a maximum daily difference of 25% below Q_i, as *ET* dominated the peatland water budget until 12 September (see Fig. 3.6). In response to the subsequent large water input on that day (Fig. 3.8a), Q_o increased at a faster rate than recorded at Q_i (Fig. 3.8b). Q_i and rainfall inputs were enough to exceed *ET* and net groundwater losses resulting in a peatland water table higher than -0.15 m (Fig. 3.8c). Of the 15 mm of water input, 26% was directed to *ET* and 53% to peat storage leading to a small delay in the release of stored water until 13 September. The rainfall event on 2 October (3 mm) produced only a small increase in the peatland water table to -0.17 m (Fig. 3.8c). Following the 7 October event, Q_o was 43% greater than at Q_i. For this event, only 8% of the 20 mm rainfall input was directed to *ET* and 15% to peat storage. Smaller rainfall events during the remainder of the study period (e.g. 23 October) prolonged the recession of the 7 October event. During this time, the peatland water table varied between -0.11 m to -0.15 m and Q_o was 29 to 43% greater than Q_i (Fig. 3.8b). Q_o was greater than Q_i when the peatland water table was higher than -0.15 m. Conversely, Q_o was less than Q_i when the water table dropped lower than -0.15 m, indicating an active runoff threshold (Fig. 3.8c).

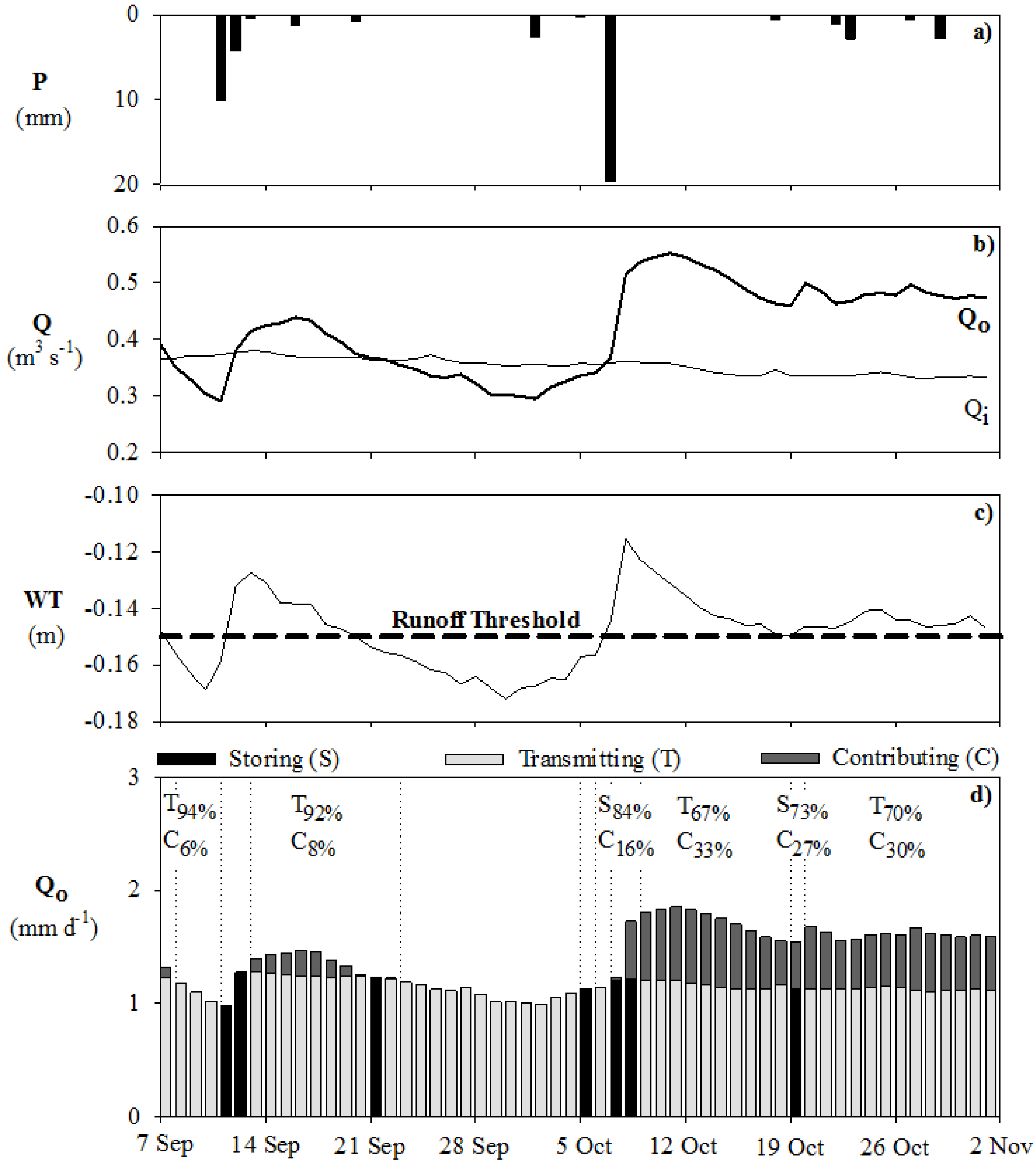


Figure 3.8: Streamflow response and peatland function during fall 2011 study period: (a) Rainfall recorded during the study period; (b) Stream hydrographs of Q_i (thin line) and Q_o (thick line); (c) Water table elevation recorded in the peatland (P6) showing the runoff threshold (-0.15 m); and, (d) Daily hydrological function of the peatland producing subsequent streamflow response. Changes in predominate function (i.e. storing or transmitting) are separated by dotted lines. Percentages associated with each predominant function (when not 100%) are the average percentages that each function provided in each phase.

The peatland predominantly transmitted water 89% of the time, and stored water the other 11% of the 56 day study period (Fig. 3.8d). The storage function was observed following large rainfall events. The active runoff threshold corresponds to a functional change in the peatland between predominantly transmitting water and contributing internally generated runoff. The peatland contributed a higher percentage of internally generated runoff following the 7 October rainfall event compared to the 11-13 September event.

3.4.2 Temporal Variation in Scaled Landscape Unit Outflows

The antecedent moisture and pattern of wetting up or drying of PFCC over the 10 year record can be anticipated to some extent by the depth to water table and difference in annual *P* and actual *ET (AET*; Fig. 3.9). Over the past decade, PFCC has had, for the most part, a moisture surplus (i.e. $P > AET$). The mature pine upland had positive mean annual ΔS (17 mm y^{-1}) over the 10 years of record associated with the rising water table (Fig. 3.9), which reflects the cumulative moisture surplus over time. The water table at the mature black spruce site (hereafter peatland, assuming similar hydrologic controls) remained several meters closer to surface than in the mature pine upland. The near zero mean annual ΔS (1 mm y^{-1}) estimated for Pine Fen over the 10 years of record reflects the smaller storage capacity and short water 'memory'. The greater disparity in the position of the water table among sites during dry years (2001-2003) compared to recent years indicates several years of annual moisture surplus were needed to fill the large storage capacity of the pine upland, whereas the peatland filled its relatively smaller storage capacity more quickly when surplus water was available (~1-2 years; Fig. 3.9).

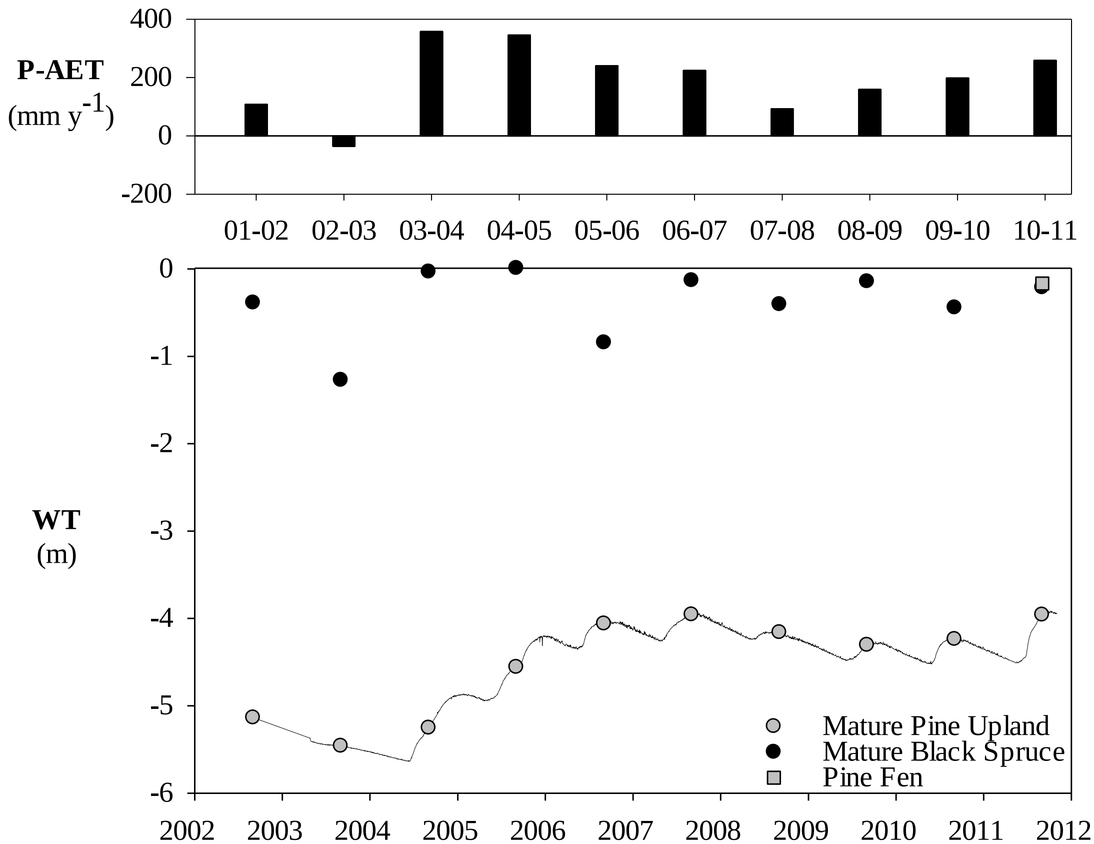


Figure 3.9: Area-weighted total annual *P*-actual *ET* (i.e. moisture availability) using the four landscape units within PFCC over 10 hydrologic years (beginning 1 October and ending 30 September of the following year; top panel). Water table depths relative to surface as of 31 August 2002-2011 at the mature jack pine upland site (OJP) and mature black spruce site (water table is assumed indicative of Pine Fen) located in a moderate topographic depression (OBS; bottom panel). The 31 August 2011 water table in Pine Fen was included for comparison. Elevations of ground surface are 461.97 m a.s.l. (OJP), 540.0 m a.s.l. (OBS), and 452.05 m a.s.l. (Pine Fen, P6).

Groundwater contribution from landscape units was influenced by the inter-annual seasonal pattern of wetting and drying reflected in lateral outflow estimates. Mature pine upland continued to provide lateral outflows even in years with a moisture deficit (2002-2003). The largest mature pine upland outflows occurred during a period of relatively low annual moisture surplus (2007-2008) with an associated negative change in storage (Fig. 3.10). The rising water table following successive moisture surplus years resulted in an increase in groundwater outflow (mean groundwater recession rate increased from 0.6 mm d^{-1} in 2002-2003 to 1.1 mm d^{-1} for 2007-2008). Regenerating pine upland outflows showed a similar temporal trend as the mature pine, but with lower rates of lateral outflows. Low antecedent conditions of the pine uplands compared to the peatland resulted in larger peatland outflows than combined pine upland outflows during the large moisture surplus years of 2004-2006 (Fig. 3.10). Similarly, large moisture surplus and wet antecedent conditions in the pine uplands (small positive ΔS) and peatland (i.e. water table close to surface) resulted in large landscape unit outflows in 2010-2011 (Fig. 3.9 and Fig. 3.10). In general, peatland outflows tracked the temporal pattern in moisture availability, meaning they were higher in wetter years (e.g. 2004-2005 and 2010-2011) and lower in drier years (e.g. 2001-2002 an 2002-2003; Fig. 3.10). In all years with lake level records, lake outflows were small compared to those from the peatland and upland stands. Scaled catchment streamflow did not systematically over- or under-estimate the few discrete measurements that exist for PFCC outflow (Fig. 3.10c). The mean absolute error between estimated and measured streamflow was 0.26 mm d^{-1}.

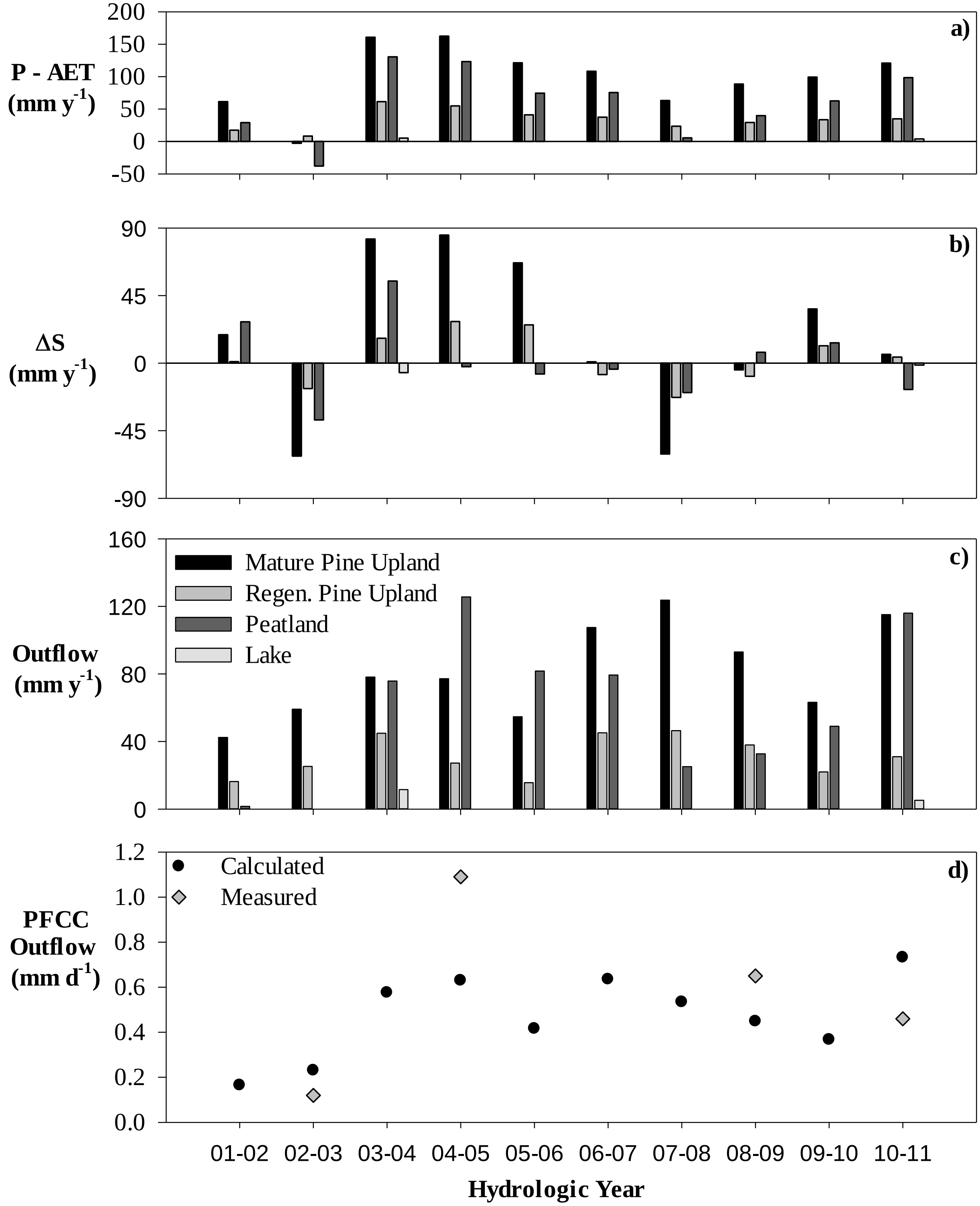


Figure 3.10: PFCC annual water balances for hydrologic years 2001-2011: a) annual *P-AET* (i.e. available moisture) for each landscape unit; and, b) calculated annual landscape unit outflows. Hydrologic years with no lake outflow indicate years with no field measurements; and, c) estimated mean daily outflows computed via scaling stand-level outflows to the catchment compared to four single-day field measurements of streamflow. Hydrologic years begin October 1 and end September 30 of the following year.

3.4.3 Water Isotope Composition

Isotopic composition of various water sources are plotted, relative to the Saskatoon, Saskatchewan meteoric water line (Fig. 3.11). Deep groundwater (i.e. thicker unsaturated zone) plotted low on the line, with average $\delta^{18}O$ and $\delta^{2}H$ values of -16.6‰ and -127.9‰. During the study period, the average stable isotope values of groundwater sourced from the peatland and shallow sand aquifer plotted close to deep groundwater. There was no significant difference between groundwater sourced from the peatland and shallow sand aquifer ($t = -0.325$, $n = 8$, $p = 0.752$). No distinct difference among streamflow, peatland, and shallow sand aquifer isotopic compositions was observed. Average $\delta^{18}O$ and $\delta^{2}H$ values at the peatland inlet (Q_{120}; -16.2‰, -124.5‰) and outlet (Q_o; -15.8‰, -121.4‰) were significantly different ($t = -5.091$, $n = 38$, $p = <0.001$). Further, Q_o showed an increased enrichment in $\delta^{18}O$ during a rainfall event (Fig. 3.12), whereas Q_{120} remained relatively stable. During this time, water collected from Pine Fen Creek ~1 km downstream of Hwy 120 (Q_S) was similar to Q_{120} (16.2‰, -125.0‰; Figure 3.11) showing little spatial variation in isotopic composition in the upstream portions of the peatland. During the intensive sampling period (7 to 20 September), mean *EC* was greater at Q_{120} (375 μs cm^{-1}) than Q_o (322 μs cm^{-1}).

Lake water plotted below the local meteoric water line, indicating enrichment by evaporation. Average $\delta^{18}O$ and $\delta^{2}H$ values for Zeden Lake were similar to that for Ispuchaw Lake. Water collected near shore versus center of the lake clustered tightly on the plot, indicating little spatial difference in the isotopic composition of the lake.

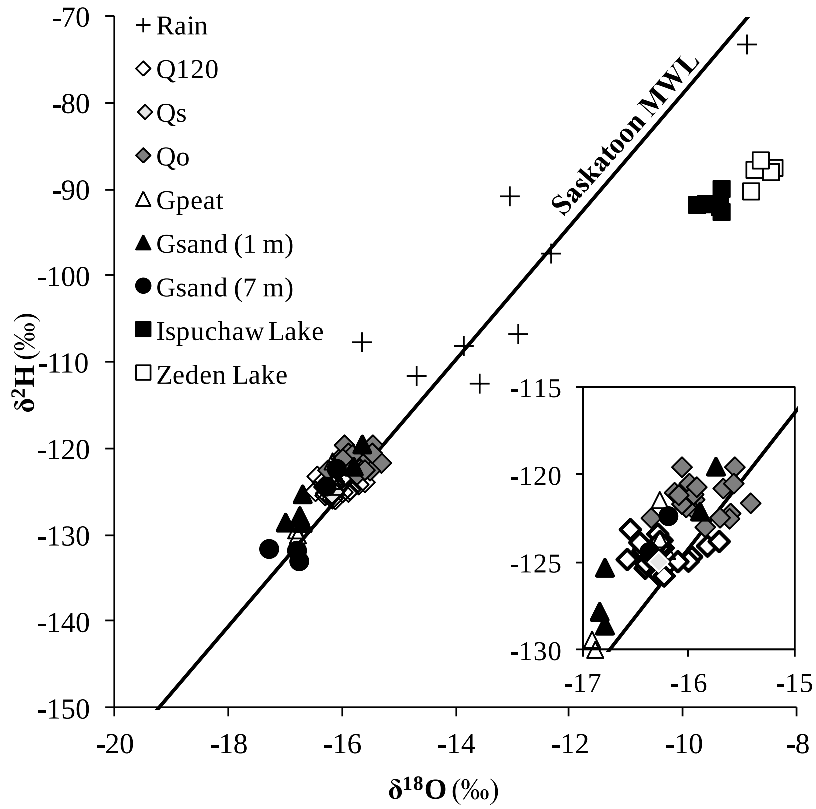


Figure 3.11: Plot of $\delta^{18}O$ and $\delta^{2}H$ showing differences in water isotope signatures for source waters in PFCC relative to the Saskatoon meteoric water line (MWL). The inset panel is a zoomed in portion of the peatland inlet (Q_{120}) and outlet (Q_o).

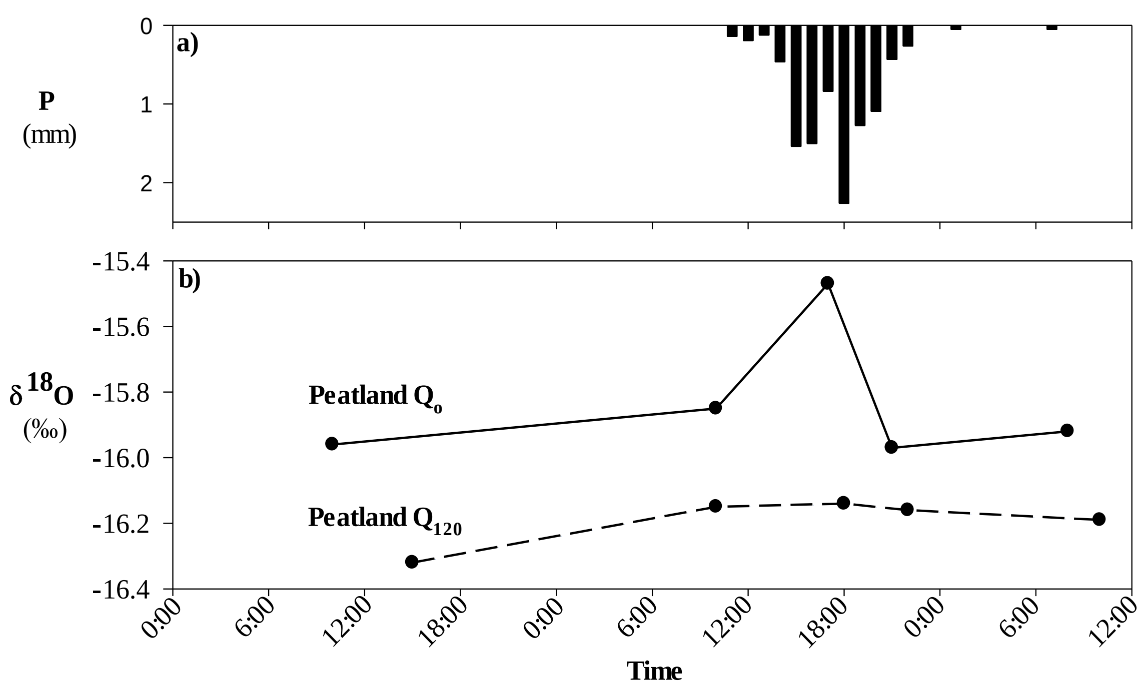


Figure 3.12: Time series of a) rainfall event and b) $\delta^{18}O$ values from peatland inlet (Q_{120}) and outlet (Q_o) during 10 September 2011 0:00 to 12 September 2011 12:00.

The isotopic signature of Pine Fen Creek flowing through glacial outwash was more depleted in $\delta^{18}O$ compared to White Gull Creek west branch in fall 2011, which flows through glacial till moraine. The same trend was observed between water sampled from Pine Fen Creek and White Gull Creek west branch in fall 2002 in the midst of the 2001-2003 regional drought, indicating there was a greater proportion of groundwater input to Pine Fen Creek (Fig. 3.13). Late winter stable isotope values collected in 2003 and 2004 indicate the composition of Pine Fen Creek was similar to deep groundwater. Pine Fen Creek was less enriched in $\delta^{18}O$ during the drought (2002) compared to the average isotope value during the wet fall of 2011.

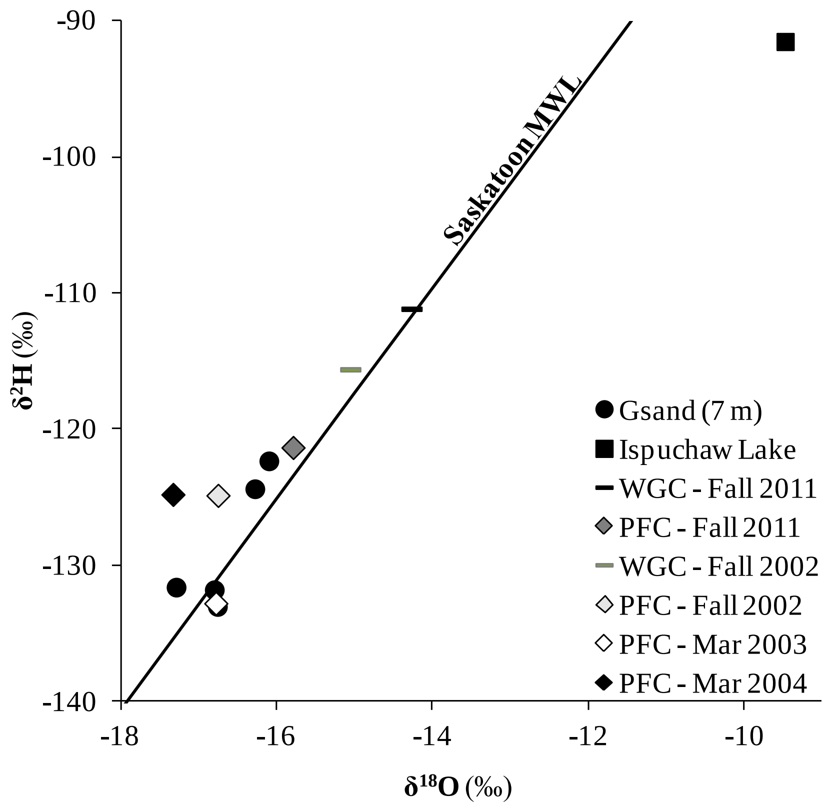


Figure 3.13: Plot of $\delta^{18}O$ and $\delta^{2}H$ showing differences in water isotope composition from sub-catchments of WGCRB with different surficial geology including White Gull Creek (WGC) west branch flowing through glacial till moraine and Pine Fen Creek (PFC) through glacial outwash.

CHAPTER 4 - DISCUSSION

The four landscape units studied herein (mature and regenerating pine uplands, lakes and peatland) had varied groundwater exchange with the underlying sand aquifer over the study period. Overall, all landscape units gained groundwater flow over the course of the study period. The pine uplands were important areas of groundwater recharge in the catchment. The lakes were a minor contributor of groundwater to the stream owing to their small size and distant location from the stream. Groundwater flowing from the pine uplands down-valley towards the stream was intercepted and delayed by the peatland due to its valley-bottom position and short water 'memory'. Results suggest that typology and topology are major controls on regulating groundwater flow to the stream in this catchment, indicating that the T^3 model (Buttle, 2006) is a useful conceptual model for explaining the hydrological processes of this catchment.

The pine uplands within PFCC provided varied recharge to the water table during the fall 2011 study period. Variation in pine upland groundwater recharge likely depends on year-to-year variations in climate, particularly annual snow accumulation, synchronicity of rainfall events compared to seasonal actual *ET*, and interception of vegetation (Smerdon et al., 2008). Annual net groundwater recharge of 103 mm y^{-1} was recorded in the mature pine upland crests of PFCC (2002-2011), which was comparable to recharge estimates for pine uplands in northern Alberta with a similar water table position (78-96 mm y^{-1}; Redding, 2009). The high permeability and infiltration capacity of the coarse-textured substrates in the pine uplands (Redding, 2009) were critical in maintaining recharge within PFCC. However, the pine uplands' water table hydrographs indicate the fall 2011 study period occurred during a time of sinusoidal phase shift between a predominantly positive flux at the water table (i.e. recharge) to a negative flux (i.e. drainage; see Fig. 3.9). In coarse-textured landscapes, the timing and magnitude of the water table response (i.e. sinusoidal phase shift) to infiltrating rainfall through the unsaturated zone will vary depending on antecedent conditions, actual *ET* demand, hydraulic properties, and depth to the water table (Smerdon et al., 2008). In a drier period, greater depth to the water table will often result in an increased lag and attenuation of the rainfall signal (Cuthbert et al., 2010). Wet antecedent moisture conditions combined with large summer precipitation preceding the study period (70% greater than the climate normal) likely exceeded actual *ET* and canopy interception to aid in groundwater recharge to the water table through the summer months (Smerdon et al., 2008). Increased net groundwater drainage in the mature pine uplands over the course of the fall study period was likely a result of relatively less fall precipitation. In contrast, net groundwater recharge measured into the fall in the regenerating pine uplands could be attributed to the younger trees transpiring and intercepting less precipitation. This would result in higher soil moisture at depth; a trend consistently reported in the literature (e.g. Elliot et al., 1998).

Groundwater originating in the pine uplands likely flowed along the sand-till interface toward the stream. Clay-rich glacial till (within Saskatchewan) has an estimated vertical hydraulic conductivity (K_v) of 10^{-11} to 10^{-12} m s^{-1} (Keller et al., 1988; Shaw and Hendry, 1998), whereas sandy soils have a typical K_v of 10^{-4} m s^{-1} (Fetter, 2001). Therefore, rainfall infiltrating the soil will preferentially move vertically to an underlying clay-rich layer where water is then redirected laterally, flowing above the confining layer (e.g. Haynes and Mitchell, 2012). The hydraulic conductivity discontinuities that create this flow mechanism are common in humid and steep environments where large K_v differences occur at soil-bedrock interfaces, a situation reported for the Boreal Shield (Peters et al., 1995; Buttle and McDonald, 2002), as well as the mountainous areas of New Zealand (McGlynn et al., 2002), and southeastern USA (Tromp van Meerveld and McDonnell, 2006). Within PFCC, lateral groundwater transmission from the pine upland towards the stream (Fig. 4.1) is likely similar to the concept proposed for a pine upland-peatland system in Alberta (Redding, 2009), which also lies in the Boreal Plain. However, the ultimate fate of the groundwater sourced from the pine uplands depends on whether its flow is interrupted by wetlands, riparian areas, or peatlands before it reaches the stream channel (Soulsby et al., 2006; Jensco et al., 2009; Spence et al., 2011).

There is no reach of Pine Fen Creek that originates in the pine uplands. Instead, the groundwater flow originating in the uplands must pass through (or beneath) the peatland before reaching the stream, as illustrated on the water table map and hydrogeological cross sections. Groundwater flow sourced from the pine uplands is transiently detained by the peatland indicated by the slower water table recession hydrographs following rainfall events. Spence et al. (2011) showed that a subarctic boreal wetland stored upland runoff near its margin and did not distribute the (subsurface) water uniformly across the peatland in a way that influenced streamflow response. Although it has been shown that elevated water tables in uplands can affect streamflow response despite the two being indirectly connected via a peatland (Branfireun and Roulet, 1998), this was not likely the case at PFCC. The results from PFCC can be explained by the large spatial extent of the peatland. Pine Fen is located in an unconfined valley with an average hillslope-stream distance of 1050 m. This is a considerably greater distance than other Boreal catchments that reported upland-peatland connectivity influenced streamflow response (Branfireun and Roulet, 1998; Kværner and Kløve, 2008). Their hillslopes were, on average, 50 m from the stream. Hillslope seepage at PFCC likely flowed laterally into the underlying sand aquifer or into the peat subsurface (Redding, 2009), but was only a minor flux to the peatland (Fig. 4.1). However, there are some areas of the catchment where groundwater originating in the upland is probably less detained in the peatland as it flows toward the stream, as the peatland has geometry where it is ~9 times narrower at its southernmost extent than its maximum width (i.e. hillslope-stream distance of 125 m). The spatial variability of hydrologic connectivity between upland-peatland-stream should be investigated in detail due to differences in hillslope-stream distances along Pine Fen. To resolve the issue of how far the stream channel needs to be from the hillslope before the hillslopes become hydrologically irrelevant requires a future detailed investigation of isotopic signatures of hillslope, peatland, and stream water.

Overall, the levels of Zeden and Ispuchaw Lakes were maintained by large net gains of groundwater from the underlying sand aquifer over the study period. Lakes situated in coarse-textured outwash have been termed 'evaporative windows' (Smerdon et al., 2005), given the fact that their largest hydrological loss is typically evaporation. Evaporation was the largest flux from the lakes in fall 2011, as corroborated by the isotopically enriched $\delta^{18}O$ and $\delta^{2}H$ values of lake water compared to the local meteoric water line. The duration and timing of evaporation was important in controlling net groundwater exchange between the lakes and the underlying sand aquifer. Although Ispuchaw (2.6 mm d^{-1}) and Zeden Lakes (1.1 mm d^{-1}) gained groundwater from the underlying aquifer during the fall study period, the over-winter water balance indicates the lakes can also recharge the underlying sand aquifer. Over the 2010-2011 winter, the lakes provided -1.4 mm d^{-1} to -1.2 mm d^{-1} of water to the sand aquifer. Aquifer discharge to PFCC lakes during fall 2011 was an average 58% less than that measured at another northern Alberta Boreal lake, intensively studied during two moisture deficit years (Smerdon et al., 2005), suggesting inter-annual climate variability may be important in controlling aquifer-lake interactions.

Although lakes within PFCC were primarily gaining in fall 2011, flow reversals between the lake and its bed were observed on short time scales. Since these flow reversals were documented with piezometers situated near the lake shore, they likely represent varying hillslope-lake interactions, rather than the whole lake basin-aquifer interaction. The small rainfall events that occurred during the period of sinusoidal phase shift in the pine uplands' water table hydrograph likely increased the variation in water table gradients between the lake level and adjacent hillslope. For example, the increased water table gradient between the adjacent hillslope and Zeden Lake level following the 11-13 September rainfall event, weakened the hydraulic gradient and led to a flow reversal. Rainfall events and the development of groundwater ridging in the hillslope has caused flow reversals to occur in other outwash lakes (Cherkauer and Zager, 1989; Smerdon et al., 2005).

The channelized Ispuchaw Lake outflow is unlikely to be an important flux to Pine Fen Creek. Mean Ispuchaw Lake outflow only made up 7% of total surface inflows to the peatland during the study period. The small amount of water contributed from the lakes to the stream was relatively minor. In addition, lake outflow was intercepted by the northeastern part of the peatland and likely integrated into subsurface flow (Smerdon et al., 2005) due to the high surface porosity of the peat. Thus lake water was not likely distributed to the main stem of Pine Fen Creek (Spence et al., 2011). This conclusion is corroborated by the dissimilar stable isotope composition of lake water and streamflow. Both surface and subsurface inflow from the lake was detained long enough in the peatland such that it took on the isotopic characteristics of peatland groundwater.

The peatland was an area of groundwater discharge from the underlying sand aquifer, except at its margins. Consistent positive groundwater gradients recorded at the peatland margins indicate hillslope water likely recharged peatland storage (Spence et al., 2011). In contrast, most of the groundwater from the underlying aquifer was forced to exfiltrate to the deeper peat layers in the southern peatland area as boreholes indicate the sand overlying clay-rich glacial till was only a few metres thick. Flow reversals did occur in one well at the peatland margin (P6) preceding rainfall events, which has been reported by others (e.g. Devito et al., 1996). VHGs measured between humified peat and underlying mineral soil were much steeper (-0.37 to 0.11 m m^{-1}) than those reported for Glacial Lake Agassiz peatlands in Minnesota (-0.04 - 0.03 m m^{-1}; Siegel et al., 1995). However, they were similar to a Boreal Shield peatland underlain by a thin sand aquifer (-0.18 to 0.15 m m^{-1}; Branfireun and Roulet, 1998). Similar water isotopes from the deeper and shallow underlying sand aquifers and groundwater within the peatland corroborate strong groundwater interactions, expected in permeable outwash landscapes (Freeze and Witherspoon, 1967), and likely maintained the peatland water table. Therefore, the interaction between the peatland and underlying sand aquifer is spatially diverse, but predictable based on surficial geology and proximity to uplands.

Calculation of net groundwater exchange in the peatland was sensitive to the accurate estimation of other water balance terms over the large peatland area. However, the estimation of peatland *ET* using the OBS tower (average 1.1 mm d^{-1}) was comparable to rates recorded in fall for both a black spruce treed peatland in northern Alberta (<1 mm d^{-1}; Petrone et al., 2008) and an aqualysed black spruce treed fen in Quebec (1.2 mm d^{-1}; Proulx-McInnis et al., 2012). Further, estimated Pine Fen storage change accounted for both unsaturated and saturated storage and included a constant value for S_y; methods that have been applied in other boreal peatlands (Metcalfe and Buttle, 1999; Spence et al., 2011). Although the S_y of peat decreases exponentially with depth (Hogan et al., 2006), the method applied herein was reasonable as the measured fluctuation of the water table was confined mainly to the upper peat layer (Metcalfe and Buttle, 1999). The spatial variations in water table position caused by microtopography (Metcalfe and

Buttle, 1999), precipitation and interception (Price et al., 1995; Barr et al., 2012), and inaccuracies in delineating the peatland surface area increased the difficulty in accurately quantifying net groundwater exchange for Pine Fen. Despite these potential inaccuracies, other field data including measured groundwater discharge, groundwater flow nets and hydrogeological cross sections suggest the peatland gained groundwater flow.

The difference in stability of the peatland surface inflow and outflow hydrographs suggests the peatland has a major hydrologic role in regulating groundwater flow to the stream. Although there were many beaver ponds near the peatland inflow, they were unlikely to have maintained the stable inflow given their distant location from the gauging site (~540 m upstream). However, beaver ponds and dams have previously been shown to flatten and delay the hydrograph runoff response (Woo and Waddington, 1990), depending on their available storage capacity (Burns and McDonnell, 1998). Hydrographs influenced by beaver activity should still respond to storm events (Nyssen et al., 2011), which did not occur in the peatlands' inflow hydrograph. A more likely explanation for the stable surface inflow is that its source is primarily groundwater flow. The isotopic composition of peatland inflow (Q_{120}) supports this conclusion in that it was similar to groundwater located within the deeper sand aquifer, the shallow sand aquifer and peat pore water. The pine uplands comprise 35% of the groundwater contributing area to peatland inflow providing a stable regional groundwater source and a resulting smooth hydrograph (Winter et al., 2001; Cuthbert, 2010). The remainder of the groundwater contributing area was peatland (including the upstream portion of Pine Fen). Groundwater-fed peatlands have been shown to diminish and delay hydrograph runoff peaks as a result of their threshold storage capacity (e.g. Quinton et al., 2003). The observed difference between the stable peatland inflow and responsive peatland outflow along with the more enriched isotopic signature of the peatland outflow suggests the valley bottom peatland (Pine Fen) influences groundwater flow to the stream.

Pine Fen primarily functioned as a transmitter of groundwater, but the magnitude and timing of event peaks at the peatland outflow were influenced by the efficiency at which the peatland releases water (i.e. internal runoff generation). Water table records for a Boreal Shield upland-peatland margin indicate that when the water table was -0.15 m at the onset of a 54 mm rainfall event, 6 times greater peak streamflow was observed than when the water table was at -0.20 m (with comparable rainfall; Branfireun and Roulet, 1998). Jager et al. (2009) showed streamflow from a boreal peatland located in Finland ceased when the water table was at -0.13 to -0.15 m during two growing seasons. Peatlands will attenuate groundwater flow at different rates as a result of declining K_h with peat depth (Kværner and Kløve, 2008). The zone of preferred K_h wherein groundwater moves most rapidly is generally confined to the upper 0.50 m of the peat column (Chason and Siegel, 1986; Siegel and Glaser, 1987), and decreases with depth owing to greater peat decomposition and higher compression (Hogan et al., 2006). The K_h of the upper surface layer was not measured in Pine Fen, but K_h values in the literature for the top 0.10 - 0.15 m of peat are 10^{-3} m s^{-1} for spruce-sphagnum dominated peatlands in northern Manitoba (Metcalfe and Buttle, 2001) and 10^{-2} to 10^{-4} m s^{-1} in Norway (Kværner and Kløve, 2008). However, the actual K_h-depth relationship could be site specific as soil pipes, found along passages adjacent to major tree roots where buried logs have decayed, could greatly increase groundwater transmission regardless of depth (Waddington, 2003; Holden, 2005; Rossi et al., 2012). The data suggest Pine Fen acts like other peatlands, and experiences functional changes depending on the position of the water table relative to the peat surface, meaning there is an active runoff threshold.

Peatland function was sensitive to changes in *P* and actual *ET*, particularly the timing and intensity of rainfall events and the subsequent water table position relative to the runoff threshold. Intermittent rainfall events when the water table was below 0.15 m were too small, such as the 3 mm event on 2 October, to replenish the storage deficit as interception could be up to 60% of small rainfall events in black spruce forests (Price et al.,1995). Preceding the two large rainfall events, the maximum peatland water table rose to within 0.13 m (11 to 13 September event) and 0.11 m (7 October event) of the peat surface. The result of both events was a greater streamflow response at the peatland outlet than the peatland inflow, indicating the predominate function of the peatland changed from a transmitter of groundwater to a generator of runoff (surface and subsurface). A distinctly different isotopic signature of event water at the peatland outlet compared to the inlet during the 11 to 13 September event corroborates with the peatland exhibiting a contributing function; a pattern also found in a sub-arctic boreal catchment (Spence et al., 2011). The enrichment of $\delta^{18}O$ at the peatland outlet suggests a shorter residence time (e.g. Soulsby and Tetzlaff, 2008) inhibiting the stream water to fully develop traits of peatland groundwater. The isotopic composition of stream water was likely to be comprised of a greater proportion of event water as opposed to pre-event water previously stored within the peatland and contributing catchment area during the fall period (Kværner and Kløve, 2006). The percentage of time the peatland internally generated runoff increased toward the end of the study period when the water table was high, owing to reduced *ET* and frequent rainfall events that constantly replenished peat storage. In general, data from the study period show that when the water table was higher than -0.15 m, the peatland contributed internally generated runoff to the stream; when lower than -0.15 m, the peatland functioned predominantly as a transmitter of water.

The fit of the two existing conceptual models for explaining hydrological processes regulating groundwater flow to streams in Boreal forest settings were assessed for PFCC. Devito et al. (2005b) conceptual model, constructed for use within the Boreal Plain of Alberta, uses a hierarchical approach to examine the controls on hydrologic processes that move groundwater through a catchment within a given region in the sequence: climate - bedrock geology - surficial geology - soil depth and type - topography and drainage network. Given that the climate and stratigraphy across PFCC is homogenous, the remaining drivers of hydrological functioning in the Devito et al. (2005b) model are essentially the same as in the T^3 model (Buttle, 2006). The difference, however, is that the T^3 conceptual model assesses the interactions among them, as opposed to assessing them sequentially to shed light on how groundwater may contribute to streamflow response.

At PFCC, both typology and topology were the critical factors regulating the peatlands' ability to intercept groundwater flow paths, regulate runoff responses, and influence catchment streamflow response. A peatland (or wetland) positioned in the valley-bottom has been demonstrated in other catchments to regulate streamflow response (Branfireun and Roulet, 1998; Soulsby et al., 2006; Kværner and Kløve, 2008; Spence et al., 2011) because its soil structure functions to conserve water and maintain the water table close to the surface. This results in a shorter water 'memory' and faster runoff response to moisture surplus conditions (Devito et al., 2012). In contrast, the large storage capacity of the pine uplands functions as a longer water 'memory' after dry periods, meaning that the uplands need several years of large moisture surplus to fill. Differences in water 'memory' and antecedent moisture conditions between landscape units vary with storage capacity and influences how landscape units within PFCC interconnect to regulate groundwater flow to streams. Thus, catchment streamflow response cannot simply be the addition of net groundwater inputs from each landscape unit. For example, Barr et al. (2012) showed that the addition of scaled landscape unit outflows estimated for WGCRB agreed with measured catchment streamflow, but this was only true over the long term; the mean annual outflows over a 10 year record were used in their case. On an annual basis, the water balance of Boreal Plains landscape units do not reset to the same value at the start of each cycle meaning the change in storage does not tend to zero (Devito et al., 2012). During fall 2011, the four landscape units within PFCC had a collective net groundwater gain of +16 mm over the 56 day study period, but -21 mm flowed out of the catchment outlet. Little groundwater flowed out of the watershed in the underlying sand aquifer. This was because the sand aquifer thinned to 3 m below the peatland; streamflow thus represented 98% of catchment outflow. These results demonstrate that streamflow response cannot be predicted from knowing groundwater inputs alone, even in catchments with a coarse-textured stratigraphy. The relatively shorter water 'memory' combined with the spatial arrangement of peatlands relative to other landscape units and stream channels are of primary importance in predicting groundwater contribution to catchment runoff.

Landscape units have 'memory' of the length and intensity of the seasonal and decadal pattern of wetting and drying, which influenced scaled stand-level outflows over 10 hydrologic years (2001-2011). The short water 'memory' of the peatland and direct connection to the stream channel likely influenced catchment streamflow response more than the pine uplands or lakes, especially in dry years. Water balance studies in Alberta's Boreal Plain suggest the inter-annual variability in the storage capacity of a non-peat wetland located in the valley-bottom position influenced the runoff-generation and streamflow production more than the forested and regenerating aspen uplands, and that this effect was magnified during dry years (Devito et al., 2005a). The storage capacity of the peatland is filled more quickly in response to short term deviations in moisture surplus relative to the pine uplands resulting in a shorter return period for runoff (Devito et al, 2012). For example, low antecedent moisture conditions of the pine uplands compared to the peatland following the dry years (2001-2003) resulted in larger peatland outflows than combined pine upland outflows during the wet years (2004-2006). In contrast, large pine upland outflows occurred after several successive moisture surplus years that filled available storage and continued to provide lateral outflows during a mesic year (2007-2008). In this same year, the peatland responded more to the short term deviation in weather by conserving water, indicated by the lowest peatland outflows since the dry years. The peatlands' soil structure (high specific yield and water holding capacity) would likely result in lateral outflows (i.e. subsurface flow) draining more slowly from the peat as the transmissivity feedback diminishes, and become a key source to low flows (Smakhtin, 2001; Kværner and Kløve, 2008). Other peat and non-peat wetlands have been shown as sources of baseflow in drier years (Roulet 1990; Devito et al., 2005a; Jager et al., 2009). Although there were few measurements of PFCC streamflow to corroborate estimates of scaled catchment streamflow, isotope samples from Pine Fen Creek were sampled in fall 2002, when the region was in the midst of a dry period (2001-2003). The highly negative stable isotopic values of stream water imply that water contributing to streamflow was resident longer in the catchment with a lower proportion of event water than the stream water isotopes collected during the wetter 2011 study period (Hayashi et al., 2004). This suggests the peatland functions primarily as a transmitter during dry years, moving deeper peat groundwater from the underlying sand aquifer to the stream (Fig. 4.1). In general, annual peatland outflows tracked the temporal pattern in moisture availability, which corroborates with the detailed measurements in fall 2011 that indicate peatland function (i.e. water table-runoff generation threshold) was sensitive to seasonal weather. Therefore, the peatland regulates groundwater contributions to the stream not only in response to individual storm events, based on detailed fall 2011 measurements, but also on a year-to-year basis, depending on antecedent moisture conditions and climate.

In summary, these results provide a conceptual understanding of how hydrologic processes are interconnected between boreal landscape units and control streamflow response in coarse-textured landscapes. Highlighted was a runoff threshold that could be used to predict peatland runoff generation during low flows. Although the peatland has a large spatial extent, results suggest few measurements are needed to predict landscape function and streamflow response in wet versus dry years in a peatland-dominated outwash catchment. A conceptual representation of how landscape units within PFCC interconnect to regulate groundwater flow to streams in a coarse-textured landscape (Fig. 4.1) extends the conceptual knowledge of groundwater exchange between individual landscape units and their underlying aquifer during the low flow period (see Fig. 2.7). As topography, typology and topology all seemed to simultaneously exert control on groundwater contributions to Pine Fen Creek, the T^3 model (Buttle, 2006) was deemed most useful for conceptualizing catchment hydrological function.

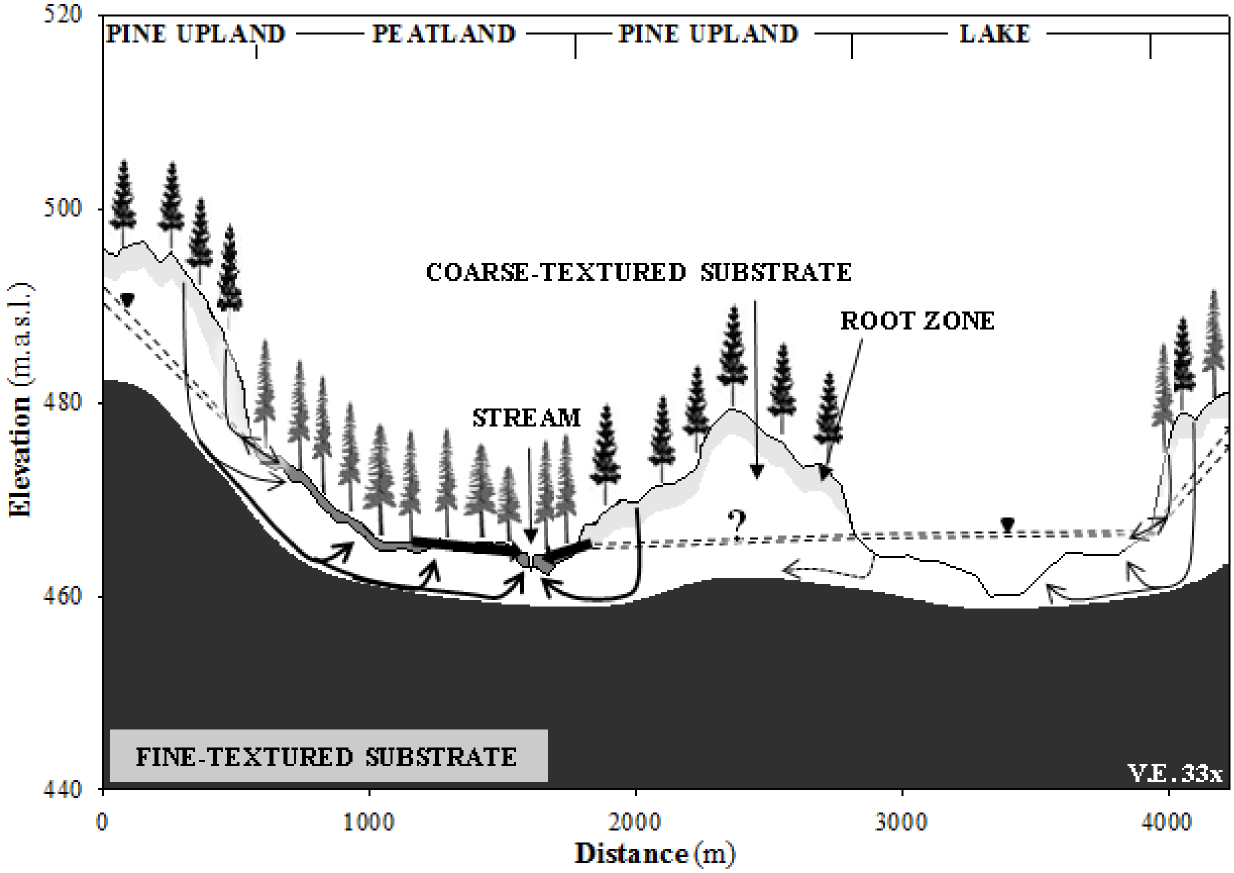


Figure 4.1: Conceptual representation of groundwater flow paths in PFCC with a range of mean water table elevations (dotted black lines) for fall 2011 (wet) and 2003 (dry), excluding the peatland. Contributing peatland function during wetter conditions would quickly release groundwater sourced nearer the stream (thick black arrow) compared to a transmitting peatland function during drier conditions, which would slowly release groundwater sourced from peat storage maintained by deeper groundwater flow paths exfiltrating to the deeper peat layers via the pine uplands (medium thick arrows). Groundwater flow paths from the underlying sand aquifer could directly contribute to the streambed (composed of mineral substrate). Other groundwater flow paths (thin black arrows) maintain lake and peatland water tables. Groundwater flow paths sourced from the lake did not likely contribute to the stream (dotted black arrow). A topographic groundwater divide may exist between the peatland and lakes; however, surficial geology and water table depths were not ground-truthed for this thesis.

CHAPTER 5 - CONCLUSIONS

5.1 Summary of Findings

Past work in the Boreal Plain has focused on the hydrological interaction between landscape units (forested uplands, lakes, and peatland) and underlying aquifers (e.g. Smerdon et al., 2005 and 2008; Redding, 2009). Although many of these studies have looked at groundwater transmission between two or three landscape units (i.e. upland to peatland or upland to lake to peatland), less known is how the typology, topology and topography (Buttle, 2006) of landscape units situated on coarse-textured substrates interact and regulate groundwater flow to streams at catchment scales. Studies in other boreal ecosystems have shown that valley-bottom peatlands can largely regulate streamflow if they are physically connected to a stream channel (e.g. Spence et al., 2011). The goal of this thesis was to improve the understanding of the roles large, valley-bottom peatlands, pine uplands and lakes have in regulating groundwater flow to Boreal Plain streams at the catchment scale.

The first objective was to characterize the hydrogeology of a typical coarse-textured catchment in the Boreal Plain. Records of borehole lithology within PFCC indicate a distribution of substrate hydraulic properties similar to glaciated outwash landscapes studied by others in Minnesota, northern Wisconsin, Nebraska, and northern Alberta. The deep glaciated terrain results in strong groundwater-surface water interaction and good connection to larger-scale groundwater flow systems, as is expected in permeable outwash landscapes (Freeze and Witherspoon, 1967). Although the delineated groundwater contributing area of Pine Fen Creek was fairly similar in size to the surface watershed, the two need not match. Regional groundwater can also help sustain baseflow in streams (Sophocleous, 2002) meaning the groundwater contributing areas in outwash landscapes can extend further past surface topographic boundaries. Thus, knowing the contributing area for groundwater within outwash catchments is critically important in properly estimating baseflow and is required for applying the water balance approach to do so.

The second objective was to measure both recharge across the water table in the pine uplands and peatland, and groundwater exchange between the four landscape units and underlying sand aquifer within the catchment (mature and regenerating pine uplands, lakes, and peatland). The pine uplands were important areas of annual groundwater recharge in the catchment. Groundwater exchange between the landscape units and sand aquifer varied over the study period. Overall, all landscape units gained groundwater flow since the study was in fall during a period of sinusoidal phase shift, as demonstrated by the pine upland water table hydrographs. The lakes acted as ‘evaporative windows’ during the study period, but changed little in level. The evaporative demand was fed by inflowing groundwater. The peatland gained water throughout the study period because it was situated in a topographic low, adjacent to the stream channel. Nearly all of the catchments’ groundwater flow was forced up into the peatland at its southernmost extent due to the thinning of the sand aquifer here. Net groundwater exchange rates estimated for each landscape unit within PFCC were similar to those observed in coarse-textured outwash landscapes with comparable climate regimes and topography (e.g. Smerdon et al., 2005; Winter et al., 2001). Results from this study thus geographically extend the understanding of landscape unit groundwater exchange across the Boreal Plain.

The third objective of the study was to evaluate how landscape units interact to regulate groundwater flows to streams. Results showed that the typology and topology were critical factors in determining how groundwater flow from landscape units (pine uplands, lakes, and peatland) interacts to regulate groundwater contributions to the stream. Groundwater flow nets and lateral hydraulic gradients indicated that the pine uplands provided constant groundwater flowing down-valley towards Pine Fen Creek. The lakes had a minor influence on regulating groundwater contribution to streamflow due to their distant location from the channel and relative small areal extent. Distinct differences in the isotopic composition of water collected from the lakes and Pine Fen Creek support this conclusion. In contrast, the peatland had a major hydrologic role in regulating groundwater contributions to the stream. The peatland intercepted and regulated groundwater originating from the pine uplands and lakes due to its valley-bottom position and short water 'memory'. The stable isotopes were not different between the composition of water within the peatland, underlying sand aquifer, and Pine Fen Creek; all tended toward the deep groundwater isotopic signature. This supported the conclusion that the peatland intercepted groundwater flows from the two other landscape units and regulated its transmission to the stream. Data also support the use of Buttle's (2006) T^3 model for predicting groundwater transmission to streams in this hydrogeological setting.

Predicted lateral outflows for the peatland over 10 years indicate there was likely year-to-year variation in peatland function, switching between a source and sink of runoff to streamflow, depending on whether there were conditions of water surplus, water deficit, or near balance. Results from fall 2011 provided further insight into the trigger for functional change, indicating it was dependent on the position of the water table relative to the peat surface. The results suggest the water table threshold for peatland runoff generation at this site is -0.15 m.

The presence of this dynamic peatland function within a single season and over the long term illustrates the sensitivity of this hydrologic system to changes in the ratio of P to ET. Slight variations in the peatland water table-runoff generation threshold may occur seasonally meaning lower runoff could occur at higher water levels (Jager et al., 2009) due to compression of the surface peat in drier summer conditions (e.g. Hogan et al., 2006). Raised water levels in spring due to frost table depth (Wright et al., 2009) may have the opposite effect. Other researchers have also found variability in hydrologic processes between landscape units and sand aquifers during other seasons (e.g. snowmelt and summer) within the Boreal Plain (e.g. Smerdon et al., 2005; 2008). The connectivity of landscape units and synchronicity of inputs (rainfall, snowmelt, and runoff) to a valley-bottom wetland influenced wetland function and catchment streamflow response in the subarctic boreal forest (Spence et al., 2011). This concept needs exploration to determine its applicability to valley-bottom peatlands in the Boreal Plain.

5.2 Implications of Thesis Findings

The storage, transmitting, and contributing properties of peatlands are not adequately represented in most hydrological models (Whitfield et al., 2009), especially at low flows (Davison and van der Kamp, 2008). The findings presented here support the approach to define the peatland as a separate hydrologic response unit (Devito et al., 2005b). Needed for modeling is an understanding of both peatland storage capacity, as classifying peatlands as swamp, marsh, or forested would not provide appropriate values for the hydraulic soil properties, and hydrogeomorphic position in the landscape. Runoff generation algorithms in hydrological models should thus account for the high storage capacity of peatlands. Whitfield et al. (2009) suggested a suitable metric for this would be the position of the water table, and results from PFCC support this notion. Further, results show catchment streamflow response during low flows was not simply the addition of net groundwater inputs from each landscape unit; therefore, routing algorithms in distributed hydrological models need to consider the hydrological (groundwater) linkages of landscape units within the catchment. Not accurately representing these linkages could result in models failing to replicate the volume and timing of catchment hydrographs as peatlands make up a high percentage of land cover in the Boreal Plain (21%).

Knowing the relationship between water table position and peatland runoff generation could also prove useful in understanding the hydrologic consequences and susceptibility of these and similar catchments to hydrologic change via climate change (Schindler and Donahue, 2006; Bergengren et al., 2011) or energy or forestry sector disturbances (Devito et al., 2011; Seitz et al., 2011). PFCC is situated near a climate-sensitive boundary; the transition between boreal forest and aspen parkland coincides with equal precipitation and potential evapotranspiration (Hogg, 1997). Changes in the regional water balance as a result of climate change could result in a transition of vegetation structure to greater grassland within the boreal forest (Bergengren et al., 2011). As peatlands are typically found in moisture positive regimes (i.e. precipitation greater than potential evapotranspiration), the peatland-dominated outwash catchment studied herein could be useful in forecasting exercises to understand changing regional water balances on hydrologic function and streamflow production in wet or dry years. Further, there is legislated requirement in Canada (*Canadian Environmental Assessment Act*) to restore natural landscapes following industrial resource extraction, and that these reconstructed ecosystems be sustainable. The PFCC could thus be used as a reference site for evaluating optimal approaches to re-establish the upland-peatland systems within post-mined landscapes as it has similar surficial geology to much of the oil sands leased for development. For example, it could be used to further understand the disparities between general models (Price et al., 2010) and natural catchments.

5.3 Future Work

This thesis describes the (subsurface) hydrological function of a typical Boreal Plain catchment with a coarse-textured substrate from a meso-scale perspective, which is critical foundational knowledge that can be used to address a number of questions at various scales. For example, a major challenge in hydrology is how to upscale peatland hydrologic processes to estimate streamflow response at regional spatial scales (Whitfield et al., 2009). These processes are important in catchments where peatlands exist and is dependent on the relative proportion of peatland contributing area and their location within the catchment. Within the ~140 000 km^2 of land leased for oil sands development in northeastern Alberta, >65% is comprised of peatlands (Price et al., 2010). At the end of 2011, the total actively disturbed footprint includes 714 km^2 of cleared and disturbed terrestrial and aquatic area (Government of Alberta, 2012). The Athabasca sedimentary formation, similar to the geological area being developed in Alberta, extends into Saskatchewan and bitumen, the petroleum form present in oil sand has been located (Government of Saskatchewan, 2010). The rapidly expanding energy sector will continue to cause substantial land disturbance. Can a peatland runoff threshold prove useful to hydrological modelling for prediction of the impact of peatland degradation and forest clearing on downstream flooding or hydrograph peaks for a larger regional area?

Many beaver dams were observed in the northern and southern reaches of Pine Fen Creek during thesis data collection. Beaver have been shown to influence hydrologic processes in peatland streams located in alluvial valleys (Janzen and Westbrook, 2011) and valley-bottom outwash (Watters and Stanley, 2007). The presence of beaver dams along stream channels within PFCC could thus influence groundwater flow paths. Beaver dams can reduce flow velocity and regulate low flows (Woo and Waddington, 1990), attenuate water table decline in drier summer months (Westbrook et al., 2006), and retain event water due to a greater pre-event storage capacity (Burns and McDonnell, 1998). Near beaver ponds, Westbrook et al. (2006) showed increased hydraulic gradient between the stream and adjacent riparian area, which increased riparian water tables and recharge to the underlying aquifer. Present dams located in PFCC (n = 29) are generally built perpendicular to the direction of groundwater flow, which could steepen the down-valley gradient (Westbrook et al., 2006). The beaver population significantly decreased by fur trapping during the European settlement of North America and their population has only recovered to ~ one-tenth of historic size (Naiman et al., 1986). Could beaver dams along Pine Fen Creek change the hydrologic function of the peatland and catchment streamflow response especially during low flows?

Peat depth measurements across Pine Fen (mean 0.65 m) indicate peat accumulation is thin relative to other peatlands. For example, mean peat depths of 1.2 m (Metcalfe and Buttle, 2001), 1.5 m (Devito et al., 2005a), and 2 m (Branfireun and Roulet, 1998; Sandhill Fen, Hogan et al., 2006) have been measured in peatlands situated in the Canadian Boreal forest. Organic matter accumulates in peatlands when net primary productivity rates exceed decomposition rates (Clymo, 1984). Decay rates are greater in the oxic surface layers of peat (i.e. acrotelm) and decrease with depth as a result of permanent water saturation (Clymo, 1984). Peat thickness increases as the litter (i.e. above and below ground surface plant remains) becomes saturated by the water table. Future climate change scenarios suggest this region may see a change in mean annual air temperature of 5.5°C and potential *ET* could exceed expected increases in *P* (Schindler and Donahue, 2006). This may potentially decrease groundwater recharge and subsequently reduce groundwater flows. The expected increase in warmer temperatures and drier conditions that lower the water table could limit peat development or increase peat decomposition (Morris et al., 2011b). Any change in peatland thickness could have important implications for water-holding capacity and thus hydrological function. Hydrological modelling could be used to predict the point functional integrity of peatlands is compromised in terms of its ability to regulate flow to streams from other landscape units. How will peatland hydrological function (and indeed their existence) in the low boreal region be influenced by climate change?

LIST OF REFERENCES

Anderson, M.P. and Munter, J.A. 1981. Seasonal reversals of groundwater flow around lakes and the relevance to stagnation points and lake budgets. *Water Resour. Res.* 17: 1139-1150

Attanayake, M.P. and Waller, D.H. 1988. Use of seepage meters in a groundwater – lake interaction study in a fractured rock basin – a case study. *Can. J. Civ. Eng*. 15: 984-989

Barr, A.G., van der Kamp, G., Black, T.A., McCaughey, J.H. and Nesic, Z. 2012. Comparing water and energy balance closure at the BERMS Flux Towers 1999-2009. *Agri. For. Meteor.* 153: 3-13

Bay, R.R. 1969. Runoff from small peatland watershed. *J. Hydrol.* 9: 90 - 102

Bergengren, J.C., Waliser, D.E. and Yung, Y.L. 2011. Ecological sensitivity: a biospheric view of climate change. *Climate Change* 107: 433-457

Black, P.E. 1997. Watershed function. *J. Am. Water. Resour. Assoc*. 33: 1-11

Boelter, D.H. 1969. Physical properties of peats as related to degree of decomposition. *Soil Sci. Soc. Am. J.* 33: 606-609

Born, S.M., Smith, S.A. and Stephenson, D.A. 1979. Hydrogeology of glacial-terrain lakes, with management and planning applications. *J. Hydrol.* 43: 7-43

Bracken, L. and Croke, J. 2007. The concept of hydrological connectivity and its contribution to understanding runoff-dominated geomorphic systems. *Hydrol. Process.* 21: 1749-1763

Branfireun, B.A. and Roulet, N.T. 1998. The baseflow and storm flow hydrology of a precambrian shield headwater peatland. *Hydrol. Process.* 12: 57-72

Burns, D. A. and McDonnell, J. J. 1998. Effects of a beaver pond on runoff processes: comparison of two headwater catchments. *J. Hydrol.* 205: 248-264

Buttle, J. M. and McDonald, D. J. 2002. Coupled vertical and lateral preferential flow on a forested slope. *Water Resour. Resear.* 38: 1060

Buttle, J.M., Dillon, P.J. and Eerkes, G.R. 2004. Hydrologic coupling of slopes, riparian zones and streams: an example from the Canadian Shield. *J. Hydrol.* 287: 161-177

Buttle, J.M. 2006. Mapping first-order controls on streamflow from drainage basins: The T3 template, *Hydrol. Process.* 20: 3415-3422

Campbell, J.E. and Simpson, M.A. 1986. Saskatchewan surficial geology terrain classification. Saskatchewan Research Council, Saskatoon. Digitized ArcGIS shapefile, 2004

Campbell, C.C., Campbell, I.D. and Hogg, E.H. 1994. Lake area variability across climatic and vegetational transect in southeastern Alberta. *Geographie physique et Quaternaire*. 48: 207-212

Chapman, T.G. 1990. Natural processes of groundwater recharge and discharge, in Groundwater and Environment, pp. C1-C21. Universiti Kebangsaan Malaysia, Kota Bahru, Malyasia

Chason, D.B. and Siegel, D.I. 1986. Hydraulic conductivity and related physical properties of peat, Lost River Peatland, Northern Minnesota. *Soil Sci.* 142: 91-99

Cherkauer, D. S. and Zager, J.P. 1989. Simulating the influence of lake position on groundwater fluxes. *J. Hydrol.* 109: 167-184

Clymo, R.S. 1984. The limits to peat bog growth. *Philosophical Transactions of the Royal Society of London B* 303: 605-654

Cuthbert, M. 2010. An improved time series approach for estimating groundwater recharge from groundwater level fluctuations. *Water Resour. Res.* 46: W09515

Cutler, P.M., Colgan, P.M. and Mickelson, D.M. 2002. Sedimentologic evidence for outburst floods from the Laurentide Ice Sheet margin in Wisconsin, USA: implications for tunnel-channel formation. *Quaternary International* 90: 23-40

Davison, B. and van der Kamp, G. 2008. Low-flows in deterministic modelling: a brief review. *Can. Water. Resour*. J. 33: 181-194

Devito, K.J., Hill, A.R. and Roulet, N. 1996. Groundwater-surface water interactions in headwater forested *Wetlands* of the Canadian Shield. *J. Hydrol.* 181: 127-147

Devito, K.J., Waddington, J.M. and Branfireun, B.A. 1997. Flow reversals in peatlands influenced by local groundwater systems. *Hydrol. Process.* 11: 103-110

Devito, K.J., Creed, I.F. and Fraser, C.J. 2005a. Controls on runoff from a partially harvested aspen-forested headwater catchment, Boreal Plain, Canada. *Hydrol. Process.* 19: 3-25

Devito, K., Creed, I, Gan, T., Mendoza, C., Petrone, R., Silins, U. and Smerdon, B. 2005b. A framework for broad-scale classification of hydrologic response units on the Boreal Plain: is topography the last thing to consider? *Hydrol. Process.* 19: 1705-1714

Devito, K., Mendoza, C. and Qualizza, C. 2012. *Conceptualizing water movement in the Boreal Plains. Implications for watershed reconstruction.*Synthesis report prepared for the Canadian Oil Sands Network for Research and Development, Environmental and Reclamation Research Group. 164 pp

Elliot, J.A., Toth, B.M., Granger, R.J. and Pomeroy, J.W. 1998. Soil moisture storage in mature and replanted sub-humid boreal forest stands. *Can. J. Soil Sci*. 78: 17-27

Fenton, M.M., Schreiner, B.T., Nielsen, E. and Pawlowicz, J.G. 1994. Quaternary geology of the western plains; in *Geological Atlas of the Western Canada Sedimentary Basin*, G.D. Mossop and I. Shetsen (comp.), Canadian Society of Petroleum Geologists and Alberta Research Council, URL <http://www.ags.gov.ab.ca/publications/wcsb_atlas/atlas.html>, [1 November 2012]

Ferone, J.M. and Devito, K.J. 2004. Shallow groundwater-surface water interactions in pond-peatland complexes along a Boreal Plains topographic gradient. *Water Resour. Res.* 292: 75-95

Fetter, C.W. 2001. *Applied Hydrogeology*. Fourth Edition. Prentice Hall: Upper Saddle River, New Jersey. 598 pp

Flanagan, L.B. and Syed, K.H. 2011. Stimulation of both photosynthesis and respiration in response to warmer and drier conditions in a boreal peatland ecosystem. *Global Change Biology* 17: 2271-2287

Fraser, C.J., Roulet, N.T., and Lafluer, M. 2001. Groundwater flow patterns in a large peatland. *J. Hydrol.*246: 142-154

Freeze, R.A. and Witherspoon, P.A. 1967. Theoretical analysis of regional groundwater flow 2. Effect of water-table configuration and subsurface permeability variation. *Water Resour. Res.* 3: 623-635

Freeze, R.A. and Cherry, J.A., 1979. *Groundwater*. Prentice-Hall Inc., Englewood Cliffs, N.J., 604 pp

Gibson, J.J., Prepas, E.E. and McEachern, P. 2002. Quantitative comparison of lake throughflow, residency, and catchment runoff using stable isotopes: modelling and results from a regional survey of Boreal lakes. *J. Hydrol.* 262: 128-144

Granger, R.J. and Hedstrom, N. 2010. Controls on open water evaporation. *Water Resour. Res.*7: 2709-2726

Granger, R.J. and Hedstrom, N. 2011. Modeling hourly rates of lake evaporation. *Water Resour. Res.* 7: 2727-2746

Hayashi, M., van der Kamp, G., and Rudolph, D.L. 1998. Water and solute transfer between a prairie wetland and adjacent upland, 1. Water balance. *J. Hydrol.* 207: 42-55

Hayashi, M., Quinton, W.L., Pietroniro, A. and Gibson, J.J. 2004. Hydrologic function of *Wetlands* in a discontinuous permafrost basin indicated by isotopic and chemical signatures. *J. Hydrol.* 296: 81-97

Haynes, K.M. and Mitchell, P.J. 2012. Inter-annual and spatial variability in hillslope runoff and mercury flux during spring snowmelt. *J. Environ. Monit.* 14: 2083

Healy, R.W. and Cook, P.G. 2002. Using groundwater levels to estimate recharge. *Hydrogeol.* 10: 91-109.

Hogan, J.M., van der Kamp, G., Barbour, S.L. and Schmidt, R. 2006. Field methods for measuring hydraulic properties of peat deposits. *Hydrol. Process.* 20: 3635-3649

Hogg, E.H. 1997. Temporal scaling of moisture and the forest-grassland boundary in western Canada. *Agri. For. Meteor*. 84: 115-122

Holden, J. 2005. Piping and woody plants in peatlands: cause or effect? *Water Resour. Res.* 41: W06009

Holden, J. 2009. Flow through macropores of different size classes in blanket peat. *J. Hydrol.* 364: 342-348

Hvorslev, M.J. 1951. Time lag and soil permeability in groundwater observations. *Waterways Experimental Station Bulletin* 36, United States Army Corps of Engineers, Mississippi

Jager, D.F., Wilmking, M., and Kukkonen, J.V. 2009. The influence of summer seasonal extremes on dissolved organic carbon export from a boreal peatland catchment: Evidence from one dry and one wet growing season. *Sci. Total. Environ.* 407: 1372-1382.

Janzen, K. and Westbrook, C. J. 2011. Hyporheic Flows Along a Channelled Peatland: Influence of Beaver Dams. *Can. Water Resour.* J. 36: 331-347

Jaquet, N.G. 1976. Ground-water and surface-water relationships in the glacial province of Northern Wisconsin - Snake Lake. Ground Water. 14. 194-199

Jensco, K., McGlynn, B., Gooseff, M., Wondzell, S., Becala, K. and Marshall, L. 2009. Hydrological connectivity between landscapes and streams: Transferring reach- and plot-scale understanding to the catchment scale. *Water Resour. Res.* 45: 1-16

Judd-Henrey, I., van der Kamp, G., Wismer, M. and Rodriguez-Prado, A. 2008. *Assessment of ground and surface water conditions in the Prince Albert Model Forest area.* Final Report. SRC Publication No. 11618-1E08. 47 pp

Keller, C.K., van der Kamp, G. and Cherry, J.A. 1988. Hydrogeology of two Saskatchewan tills, Fractures, bulk permeability, and spatial variability of downward flow. *J. Hydrol.* 101: 97 - 121

Kishel, H.F. and Gerla, P.J. 2002. Characteristics of preferential flow and groundwater discharge to Shingobee Lake, Minnesota, USA. *Hydrol. Process.* 16: 1921-1934

Krabbenhoft, D.P., Bowser, C.J., Anderson, M.P, and Valley, J.W. 1990. Estimating groundwater exchange with lakes: 1. the stable isotope mass balance method. *Water Resour. Res.* 26: 2445-2453

Kværner, J. and Kløve, B. 2006. Tracing sources of summer streamflow in boreal headwaters using isotopic signatures and water geochemical components. *J. Hydrol.* 331: 186-204

Kværner, J. and Kløve, B. 2008. Generation and regulation of summer runoff in a boreal flat fen. *J. Hydrol.* 360: 15-30

LaBaugh, J.W., Rosenberry, D.O. and Winter, T.C. 1995. Groundwater contribution to the water and chemical budgets of Williams Lake, Minnesota, 1980-1991. *Can. J. Fish. Aquat. Sci.* 52: 754-767

LaBaugh, J.W., Winter, T.C., Rosenberry, D.O., Schuster, P.F., Reddy, M.M. and Aitken, G.R. 1997. Hydrological and chemical estimates of the water balance of a closed-basin lake in north central Minnesota. *Water Resour. Res.* 33: 2799-2812

Lane, RJ. 1999. *Principles of Discharge Measurement*. The Water Survey of Canada. Environment Canada. 23 pp

Lin, H., Bouma, J., Wilding, L.P, Richardson, J.L, Kutilek, M., and Nielson, D.R. 2005. Advances in Hydropedology. *Advances in Agronomy* 85: 1–89

Lis, G., Wassenarr, L.I. and Hendry, M.J. 2008. High precision laser spectroscopy D/H and 18O/16O measurements of microliter natural water samples. *Analytical Chemistry* 80: 287–293

McDonnell, J.J. 1990 A rationale for old water discharge through macropores in a steep, humid catchment. *Water Resour. Res.* 26: 2821–2832

McDonnell , J. J., Sivapalan, M., Vaché, K., Dunn, S., Grant, G., Haggerty R. and Weiler M. 2007. Moving beyond heterogeneity and process complexity: A new vision for watershed hydrology. *Water Resour. Res.* 43: W07301

McDonnell , J.J., McGuire, K., Aggarwal, P., Beven, K.J., Biondi, D., Destouni, G., Dunn, S., James, A., Kirchner, J., Kraft, P., Lyon, S., Maloszewski, P., Newman, B., Pfister, L., Rinaldo, A., Sayama, T., Seibert, J., Solomon, K., Soulsby, C., Stewart, M., Tetzlaff, D., Tobin, C., Troch, P., Weiler, M., Western, A., Worman, A., and Wrede S. 2010. How old is streamwater? Open questions in catchment transit time conceptualization, modelling and analysis. *Hydrol. Process.* 24: 1745-1754

McGlynn, B. L., McDonnell, J. J. and Brammer, D. D. 2002. A review of the evolving perceptual model of hillslope flowpaths at the Maimai catchments, New Zealand. *J. Hydrol. 257:* 1-26

McGuire, K. J., McDonnell, J. J., Weiler, M., Kendall, C., McGlynn, B. L., Welker, J. M. and Seibert, J. 2005. The role of topography on catchment-scale water residence time. *Water Resour. Res.* 41: W05002

McGuire, K. J. and McDonnell, J. J. 2006. A review and evaluation of catchment transit time modeling. *J. Hydrol.* 330: 543-563

McGuire, K. J. and McDonnell, J. J. 2010. Hydrological connectivity of hillslopes and streams: Characteristic time scales and nonlinearities. *Water Resour. Res.* 46: W10543

Metcalfe, R.A. and Buttle, J.M. 1999. Semi-distributed water balance dynamics in a small boreal forest basin. *J. Hydrol.* 226: 66-87

Metcalfe, R.A. and Buttle, J.M. 2001. Soil partitioning and surface store controls on spring runoff from a boreal forest peatland basin in north-central Manitoba. *Can. Hydrol. Process.* 15: 2305-2324

Morris, P.J., Belyea, L.R. and Baird, A.J. 2011a. Ecohydrological feedbacks in peatland development: a theoretical modelling study. *J. Ecol.* 99: 1190-1201

Morris, P.J., Waddington, J.M., Benscoter, B.W. and Turetsky, M.R. 2011b. Conceptual frameworks in peatland ecohydrology: looking beyond the two-layered (acrotelm-catotelm) model. *Ecohydrol.* 4: 1-11

Naiman, R.J., Melillo, J.M. and Hobbie, J.E. 1986. Ecosystem Alteration of Boreal Forest Streams by Beaver (Castor Canadensis). *Ecology* 67: 1254-1269

National Wetlands Working Group, 1988. *Wetlands of Canada, Ecological Land Classification Series* No. 24, Ottawa

Peters, D.L., Buttle, J.M., Taylor, C.H. and LaZerte, B.D. 1995. Runoff production in a forested, shallow soil, Canadian Shield basin. *Water Resour. Res.* 31: 1291–1304

Petrone, R.M., Silins, U., and Devito, K.J. 2008. Dynamics of evapotranspiration from a riparian pond complex in the Western Boreal Forest, Alberta, Canada. *Hydrol. Process.* 21: 1391-1401

Pfannkuch, H.O. and Winter, T.C. 1984. Effect of anisotropy and groundwater system geometry on seepage through lakebeds, 1. Analog and dimensional analysis. *J. Hydrol.* 75: 213-237

Phillips, R.W., Spence, C. and Pomeroy, J.W. 2011. Connectivity and runoff dynamics in heterogeneous basin. *Hydrol. Process.* 25: 3061-3075

Poff, L.N., Allan, J.D., Bain, M.B., Karr, J.R., Prestegaard, K.L., Richter, B.D., Sparks, R.E. and Stromberg, J.C. 1997. The natural flow regime. *Bioscience* 47: 769-786

Pomeroy, J.W. and Gray, D.M. 1995. Snow Accumulation, Relocation and Management. National Hydrology Research Institute Science Report No. 7, Environment Canada: Saskatoon, 144 pp

Price, J.S. and Fitzgibbon, J.E. 1987. Groundwater storage – streamflow relations during winter in a subarctic wetland, Saskatchewan. *Can. J. Earth Sci.* 24: 2074-2081

Price, A.G., Dunham, K., Carleton, T. and Band, L. 1995. Water fluxes through the canopy and moss floor of a black spruce (*Picea mariana*) stand. *EOS supplement* 76, S118

Price, J.S. and Schlotzhauer, S.M. 1999. Importance of shrinkage and compression in determining water storage changes in peat: the case of a mined peatland. *Hydrol. Process.* 13: 2591-2601

Price, J.S. 2003. The role and character of seasonal peat soil deformation on the hydrology of undisturbed and cutover peatlands. *Water Resour. Res.* 39: WR001302

Price, J.S., McLaren, R.G. and Rudolph, D.L. 2010. Landscape restoration after oil sands mining: conceptual design and hydrological modeling for fen reconstruction. *Int. J. of Min., Reclaim. Environ.* 24: 109 - 123

Proulx-McInnis, S., St-Hilaire, A., Rousseau, A.N., Jutras, S., Carrer, G. and Levrel, G. 2012. Seasonal and monthly hydrological budgets of a fen-dominated forested watershed, James Bay region, Quebec. *Hydrol. Process.* DOI: 10.10002/hyp.9241

Quinton, W.L. and Roulet, N.T. 1998. Spring and summer runoff hydrology of a subarctic patterned wetland. *Arctic Alpine Res*. 30: 285-294

Quinton, W., Hayashi, M. and Pietroniro, A. 2003. Connectivity and storage functions of channel fens and flat bogs in northern basins. *Hydrol. Process.* 17: 3665-3684

Redding, T. 2009. Hydrology of forested hillslopes on the Boreal Plain, Alberta, Canada. Ph.D. thesis. University of Alberta. 201 pp

Redding, T. and Devito, K. 2010. Mechanisms and pathways of lateral flow on aspen-forested, Luvisolic soils, Western Boreal Plains, Alberta, Canada. *Hydrol. Process.* 24: 2995-3010

Reeve, A.S., Siegel, D.I. and Glaser, P.H. 2000. Simulating vertical flow in a large peatland. *J. Hydrol.* 227: 207-217

Rodhe A. 1987. The origin of streamwater traced by Oxygen-18. PhD thesis, Uppsala University, Uppsala

Romanowicz, E., Siegel, D. and Glaser, P. 1993. Hydraulic reversals and episodic methane emissions during drought cycles in mires. *Geology* 21, 231-234

Rosa, E. and Larocque, M. 2008. Investigating peat hydrological properties using field and laboratory methods: application to the Lanoraie peatland complex (southern Quebec, Canada. *Hydrol. Process.* 22: 1866-1875

Rosenberry, D.O. and Winter, T.C. 1997. Dynamics of water-table fluctuations in an upland between two prairie-pothole *Wetlands* in North Dakota. *J. Hydrol.* 191: 266-289

Rossi, P.M., Ala-aho, P., Ronkanen and Kløve, A.K. 2012. Groundwater-surface water interaction between an esker aquifer and a drained fen. *J. Hydrol.* 432: 52-60

Roulet, N.T. 1990a. Hydrology of a headwater basin wetland: groundwater discharge and wetland maintenance. Hydrological Processes 4, 387–400

Rouse, W.R. 2000. The energy and water balance of high-latitude *Wetlands*: controls and extrapolation. *Global Change Biol*. 6: 59-68

Saxton, K.E., Rawls, W.J., Romberger, J.S. and Papendick, R.I. 1986. Estimating generalized soil-water characteristics from texture. *Soil Sci. Soc. Am. J.* 50:1031-1036

Schafran, G.C. and Driscoll, C.T. 1993. Flow path-composition relationships for groundwater entering an acidic lake. *Water Resour. Res.* 29: 145-154

Schindler, D. W. and Donahue, W. F. 2006. An impending water crisis in Canada's western prairie provinces. *Proceedings of the National Academy of Sciences* 103: 7210-7216

Schmidt, R. *Personal communications*. Environment Canada. September 2010

Seitz, N. E., Westbrook, C. J. and Noble, B. F. 2011. Bringing science into river systems cumulative effects assessment practice. *Environ. Impact Assessment Review* 31: 172-179

Shaw, R.D. and Prepas, E.E. 1990. Groundwater-lake interactions: I. Accuracy of seepage meter estimates of lake seepage. *J. Hydrol.* 119: 105-120

Shaw, R.J. and Hendry, M.J. 1998. Hydrogeology of a thick clay till and Cretaceous clay sequence, Saskatchewan, Canada. *Can. Geotech. J.* 35: 1041-1052

Shaw, D. A., van der Kamp, G., Conly, F. M., Pietroniro, A. and Martz, L. 2012. The fill–spill hydrology of prairie wetland complexes during drought and deluge. *Hydrol. Process.* Online.

Siegel, D.I. and Glaser, P.H. 1987. Groundwater flow in a bog-fen complex, Lost River Peatland, Northern Minnesota. *J. Ecol.* 75: 743-754

Siegel. D.I., Reeve, A.S. and Glaser, P.H. 1995. Climate-driven flushing of pore water in peatlands. *Nature* 375: 531-533

Sjogren, D.B., Fisher, T.G., Taylor, L.D., Jol, H.M. and Munro-Stasiuk, M.J. 2002. Incipient tunnel channels. *Quatern. Int.* 90: 41-56

Smakhtin, V.U. 2001. Low flow hydrology: a review. *J. Hydrol.* 240: 147-186

Smerdon, B.D., K.J. Devito, and Mendoza, C.A. 2005. Interaction of groundwater and shallow lakes on outwash sediments in the sub-humid Boreal Plains of Canada. *J. Hydrol.* 314: 246-262

Smerdon, B.D., C.A. Mendoza and Devito, K.J. 2008. Influence of subhumid climate and water table depth on groundwater recharge in shallow outwash aquifers. *Water Resour. Res.*44: 1-15

Smith, A.J. and Townley, L.R. 2002. Influence of regional setting on the interaction between shallow lakes and aquifers. *Water Resour. Res.* 38. 1170-1180

Sophocleous, M. 2002. Interactions between groundwater and surface water: the state of science. *Hydrogeo. J.* 10: 52-67

Soulsby, C., Tetzlaff, D., Dunn, S. M. and Waldron, S. 2006. Scaling up and out in runoff process understanding: insights from nested experimental catchment studies. *Hydrol. Process.* 20: 2461-2465

Soulsby, C. and Tetzlaff, D. 2008. Towards simple approaches for mean residence time estimation in ungauged basins using tracers and soil distributions. *J. Hydrol.* 363: 60-74

Spence, C. and Woo, M-K. 2003. Hydrology of subarctic Canadian shield: soil-filled valleys. *J. Hydrol.* 279: 151-166

Spence, C. 2006. Hydrological processes and stream flow in a lake dominated watercourse. *Hydrol. Process.* 20: 3665-3681

Spence, C. 2007. On the relation between dynamic storage and runoff: A discussion on thresholds, efficiency, and function. *Water Resour. Res.* 43: 1-11

Spence, C., Guan. X.J. and Phillips, R. 2011. The hydrological functions of a boreal wetland. *Wetlands* 31: 75-85

Todd, A.K., Buttle, J.M. and Taylor, C.J. 2006. Hydrologic dynamics and linkages in a wetland-dominated basin. *J. Hydrol.* 319: 15-35

Tòth, J. 1963. A theoretical analysis of ground water flow in small drainage basins. *J. Geophys. Res.* 68: 4795-4812

Tromp van Meerveld H.J. and McDonnell, J.J. 2006. Threshold relations in subsurface stormflow: 1. A 147-storm analysis of the Panola Hillslope. *Water Resour. Res.* 42: W02410

van der Kamp, G. *Personal communications on water level 'plopper' specifications.* Environment Canada. September 2010

van der Vegt P., Janszen A. and Moscariello, A. 2012. Tunnel valleys: current knowledge and future perspectives. Geological Society, London, Special Publications 368

Verry, E.S., Brooks, K.N. and Barten, P.K. 1988. Streamflow response from an ombrotrophic mire, in *Proceedings of the International Symposium on the Hydrology of Wetlands in Temperate and Cold Regions*, Vol 1. The Academy of Finland, Helsinki, pp 52-59

Vidon, P. G. and Hill, A. R. 2004. Landscape controls on the hydrology of stream riparian zones. *J. Hydrol.* 292: 210-228

Vitt, D.H., Halsey, L.A., Thormann, M.N. and Martin, T. 2002. *Peatland Inventory of Alberta. Sustainable Forest Management Network*, University of Alberta. ISBN 1-55261-006-3

Waddington, J.M., Roulet, N.T. and Hill, A.R. 1993. Runoff mechanisms in a forested groundwater discharge wetland. *J. Hydrol.* 147, 37-60

Watters, J.R. and Stanley, E.H. 2007. Stream channels in peatlands: The role of biological processes in controlling channel form. *Geomorph.* 89: 97-110

Weiler M. and McDonnell J.J. 2004. Virtual experiments: a new approach for improving process conceptualization in hillslope hydrology. *J. Hydrol.* 285: 3-18

Welsh, B., Kerkhoven, E. and van der Kamp, G. 2012. A winter water balance approach to quantifying lake-groundwater interactions in the Beaver River basin, Alberta.

Proceedings of the 39th Congress of the International Association of Hydrogeologists, Niagara Falls, Ontario, September 16-21, 2012

Westbrook, C.J., Cooper, D.J. and Baker, B.W. 2006. Beaver dams and overbank floods influence groundwater-surface water interactions of a Rocky Mountain riparian area. *Water Resour. Res.* 42: W06404

Whitfield, P.H., A. St-Hilaire and van der Kamp, G. 2009. Improving hydrologic predictions in peatlands. *Can. Water Resour.* J. 34: 467-478

Whittington, P.N. and Price, J.S. 2006. The effects of water table draw-down (as a surrogate for climate change) on the hydrology of a patterned fen peatland near Quebec City, Quebec. *Hydrol. Process.* 20: 3589-3600

Winter, T.C. 1981. Uncertainties in estimating the water balance of lakes. *Water Resour. Bull.* 17: 83-115

Winter, T.C. 1986. Effect of ground-water recharge on configuration of the water table beneath sand dunes and on seepage in lakes in the sandhills of Nebraska, U.S.A. *J. Hydrol.* 86: 221-237

Winter, T.C. 2000. The vulnerability of *Wetlands* to climate change: a hydrologic landscape perspective. J. Amer. Water Resour. Ass. 36: 305-311

Winter, T.C., Rosenberry, D.O., Buso, D.C. and Merk, D.A. 2001. Water source to four U.S. *Wetlands*: implications for wetland management. *Wetlands* 21: 462-473

Wright, N., Hayashi, M., Quinton, L. 2009. Spatial and temporal variations in active layer thawing and their implications on runoff generation in peat-covered permafrost terrain. *Water Resour. Res.* 45: W05414

Woo, M. K. and Waddington, J. M. 1990. Effects of beaver dams on subarctic wetland hydrology. *Arctic*: 223-230

APPENDIX A - BOREHOLE LITHOLOGY

Table A.1: Lithology and location of geological boreholes drilled within and near PFCC.

Borehole ID	Date (2011)	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Depth (m)	Lithology Description
PFEN-OUT-01	24-Aug	520838	5971651	435.80	2.4	roadbed materials
					2.7	peat
Location Description					4.2	fine sand
Edge of Harding Rd. and east of Pine Fen surface outflow					4.4	sand with pebbles
					6.1	coarse sand
					7.0	till

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Depth (m)	Lithology Description
PFEN-OUT-02	24-Aug	520781	5971648	435.87	1.8	roadbed materials
					4.6	peaty sand
Location Description					5.3	sand
Edge of Harding Rd. and west of Pine Fen surface outflow					6.5	till

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Depth (m)	Lithology Description
PFEN-OUT-03	24-Aug	520781	5971648	435.87	1.8	roadbed materials
					2.7	medium sand w/ pebbles
Location Description					5.5	till
~2 m away from PFEN-OUT-3 borehole location.						

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Depth (m)	Lithology Description
EPUP-04	24-Aug	519128	5974444	435.61	18.3	oxidized coarse sand
Location Description						
Near east shallow pine upland piezometer nest						

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
POJR-05	24-Aug	520347	5974078	460.40	11.3	coarse sand
					12.5	unoxidized sand w/ clay
Location Description					16.8	clay
Near the old jack pine tower road piezometer						

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
SZEDN-06	25-Aug	522504	5978566	466.13	1.2	very coarse sand
					7.6	till
Location Description						
Along logging road south of Zeden Lake						

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
ISPW-07	25-Aug	521222	5983320	470.24	10.1	coarse sand
					11.6	clay
Location Description					13.1	till
Shore of Ispuchaw Lake near boat launch						

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
ISPW-08	25-Aug	521229	5983463	478.27	6.1	coarse sand w/pebbles
Location Description						
Edge of road to Ispuchaw Lake						

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
PFEN-IN-09	25-Aug	519374	5983519	468.03	0.3	peat
					0.6	stones
Location Description					1.8	coarse sand
East of Pine Fen surface inflow and south of Hwy 120					3.6	unoxidized sand w/ clay
					4.5	till

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
PFEN-IN-10	25-Aug	519132	5983285	465.61	1.8	roadbed materials
					9.1	coarse unoxidized sand
Location Description						
West of Pine Fen surface inflow and south of Hwy 120						

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
WPHUP-11	16-Sep	516478	5974183	470.27	1.5	coarse oxidized sand
					13.4	coarse unoxidized sand
Location Description					14.0	clay
Near deep piezometer on upland west of Pine Fen					15.5	till

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
PFEN-IN-12	16-Sep	519137	5983284	465.34	2.1	roadbed materials
					3.7	coarse oxidized sand
Location Description					6.7	coarse unoxidized sand
West of Pine Fen surface inflow and south of Hwy 120					9.8	refusal of drill - till

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
PFEN-IN-13	16-Sep	519137	5983284	465.34	2.1	roadbed materials
					8.8	coarse unoxidized sand
Location Description					9.4	gravel
~2 m away from PFEN-IN-13 borehole location.					10.6	refusal of drill - till

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
SEPFCC-14	16-Sep	522260	5971975	452.92	2.1	medium oxidized sand
					3.0	oxidized clay
Location Description					3.4	unoxidized clay
SE corner of PFCC near HWY 106 & Harding Rd intersect					3.7	refusal of drill - till

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
WWDR 211941		522744	5989448	450.46	0.3	topsoil
					2.4	sand
Location Description					69.5	till
Near Caribou Creek					76.2	sand

Borehole ID	Date	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m a.s.l)	Lithology Depth (m)	Lithology Description
WWDR 25171		526982	5957093	428.41	11.6	sand
					40.2	till
Location Description					43.3	sand
Near Sandhill Fen					52.7	till
					54.3	sand
					55.2	silt
					79.9	till

APPENDIX B - PIEZOMETER SPECIFICATIONS

Table B.1: Specifications of piezometers and observation wells completed in the peatland and lakes of PFCC.

Unit	Piezometer ID	Install Date (2011)	UTM (13U)		Surface Elev. (m.a.s.l)	Install Depth (m)	I.D (cm)	Length Screen (m)	Depth Top Screen (m)	Completed In	Mean WL[1] (m.a.s.l)	K[2] ($m\ s^{-1}$)	Comments
			Easting	Northing									
Peatland	WPFEN-W	16-Aug	517078	5974254	455.88	1.14	2.5	0.99	0.10	peat	455.72	2.9E-05	
	WPFEN-SP	16-Aug	517078	5974254	455.88	1.32	2.5	0.10	0.56	peat		9.4E-06	
	WPFEN-DP	16-Aug	517078	5974254	455.88	1.36	2.5	0.10	1.20	sand		8.7E-06	
	EPFEN-W	16-Aug	519027	5974454	452.05	1.13	2.5	0.98	0.09	peat	451.90	2.6E-05	
	EPFEN-SP	16-Aug	519027	5974454	452.05	1.38	2.5	0.10	0.62	peat		2.9E-05	
	EPFEN-DP	16-Aug	519027	5974454	452.05	1.39	2.5	0.10	1.23	sand		6.3E-06	
	SPFEN-W	16-Aug	520657	5971667	437.53	1.18	2.5	0.99	0.14	peat	437.35	2.0E-05	
	SPFEN-SP	16-Aug	520657	5971667	437.53	1.51	2.5	0.10	0.75	peat		1.6E-05	
	SPFEN-DP	16-Aug	520657	5971667	437.53	1.41	2.5	0.10	1.24	sand		7.5E-06	
	NPFEN-W	17-Aug	518814	5982361	469.72	1.05	2.5	0.99	0.02	peat	469.30	3.2E-05	
	NPFEN-SP	17-Aug	518814	5982361	469.72	1.09	2.5	0.10	0.33	peat		4.0E-05	
	NPFEN-DP	17-Aug	518814	5982361	469.72	1.31	2.5	0.10	1.15	sand		-	filled with peat
Lakes	ISPW-SP	25-Jun	521210	5983292	467.81	0.82	3.2	0.10	0.18	sand		1.7E-04	removed 2-Nov
	ISPW-DP	25-Jun	521210	5983292	467.77	0.79	3.2	0.10	0.65	sand		3.6E-04	removed 2-Nov
	ZEDN-SP	25-Jun	521780	5982080	467.26	0.85	3.2	0.10	0.22	sand		2.4E-04	removed 2-Nov
	ZEDN-DP	25-Jun	521780	5982080	467.30	0.76	3.2	0.10	0.63	sand		1.6E-04	removed 2-Nov

[1] mean water level measured during 7 September to 1 November 2011.
[2] saturated hydraulic conductivity estimated using falling head slug tests and Hvorslev (1951) method.

Table B.2: Specifications of piezometers and observation wells completed in pine upland within and near PFCC.

Unit	Piezometer ID	Install Date (2011)	UTM (13U) Easting	UTM (13U) Northing	Surface Elev. (m.a.s.l)	Install Depth (m)	I.D. (cm)	Length Screen (m)	Depth Top Screen (m)	Completed In	Mean WL[1] (m.a.s.l)	K[2] ($m\ s^{-1}$)	Comments
Shallow Pine Upland (toe slope)	WPUP-W	11-Aug	517039	5974264	457.24	1.27	2.5	1.13	0.06	sand	456.62	1.2E-05	
	WPUP-SP	11-Aug	517039	5974264	457.24	1.43	2.5	0.10	0.75	sand		4.4E-05	
	WPUP-DP	11-Aug	517039	5974264	457.24	1.24	2.5	0.10	1.06	sand		2.9E-05	
	EPUP-W	09-Aug	519103	5974453	452.90	1.41	2.5	1.26	0.04	sand	452.23	5.2E-06	
	EPUP-SP	09-Aug	519103	5974453	452.90	1.53	2.5	0.10	0.84	sand		1.4E-05	
	EPUP-DP	09-Aug	519103	5974453	452.90	1.53	2.5	0.10	1.35	sand		2.0E-05	
	NPUP-W	10-Aug	518962	5983612	465.78	1.27	2.5	1.15	0.02	sand	465.23	1.8E-05	fill w/sand
	NPUP-SP	10-Aug	518962	5983612	465.78	1.61	2.5	0.10	0.92	sand		8.3E-06	small clay lenses
	NPUP-DP	10-Aug	518962	5983612	465.78	1.57	2.5	0.10	1.38	sand		8.1E-05	small clay lenses
Deep Pine Upland (crest)	WPHUP	25-Aug	516476	5974172	470.46	9.12	5.1	1.50	7.68	sand	464.32	1.4E-04	
	POJT	Dec-02	520252	5974243	462.68	10.91	5.1	1.50	9.21	sand	455.38	8.5E-05	
	POJR	Dec-02	520347	5974091	460.61	9.45	5.1	1.50	7.75	sand	453.82	1.7E-04	
	POJP	Dec-02	520429	5974176	460.97	10.48	5.1	1.50	8.78	sand	455.32	1.3E-04	
	POJF	Dec-02	521126	5974371	457.29	5.84	5.1	1.50	4.14	sand	457.29		
	H02	Dec-02	523062	5977459	466.64	9.27	5.1	1.50	7.57	sand	460.40	8.4E-05	
	H94	Dec-02	522637	5973441	464.40	11.33	5.1	1.50	8.13	sand	455.77		
	H75	Dec-02	523380	5969769	444.48	10.09	5.1	1.50	8.39	sand	438.92		

[1] mean water level measured during 7 September to 1 November 2011.
[2] saturated hydraulic conductivity estimated using falling head slug tests and Hvorslev (1951) method. Both falling and rising slug tests were completed in crest upland wells.

APPENDIX C - RATING CURVES

C.1 Peatland Inflow

Overbank flooding occurred during the study period (Fig. C.1) when water reached bankfull stage (0.65 m or 462.03 m a.s.l.); therefore, velocity across the floodplain was taken to be equal to the channel velocity measurement at the bank (0.01 m s^{-1}) to produce the peatland inflow rating curve (Fig. C.2).

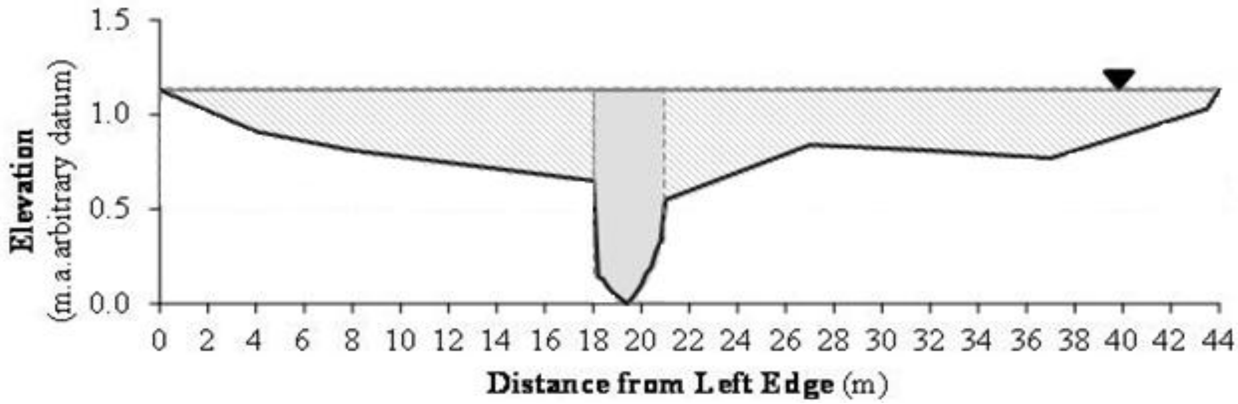


Figure C.1: Stream gauging cross-section at Pine Fen Creek surface inflow. Stream discharge was calculated using Price AA Type meter for cross-sectional area in solid grey. The cross-section area across the floodplain (grey hatched) was estimated.

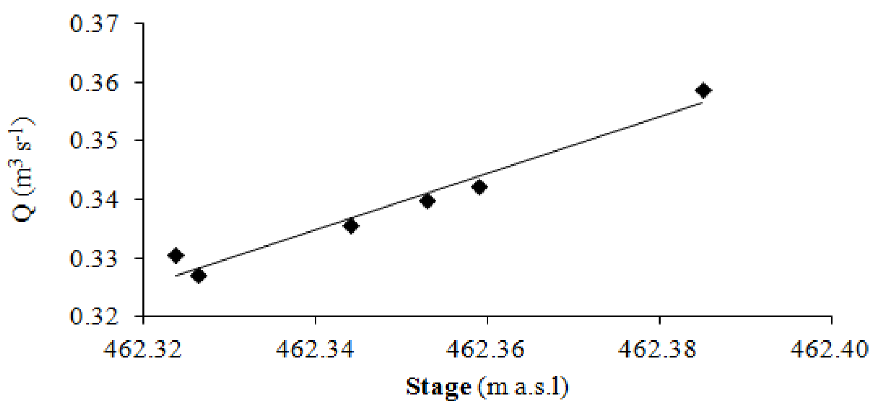


Figure C.2: Pine Fen Creek inflow rating curve for stages > 462.32 m a.s.l. with r^2 = 0.96, n = 6, P = <0.001 and Q = -221.4648 + 0.4797 • stage. The floodplain produced 34% of the total average discharge during the study period.

Estimates of velocity across the floodplain were also estimated using a Manning's n of 0.17 for flow through a sedge channel fen with a water level 0.5 to 1 m above surface (Quinton et al., 2003). Velocity estimates ranged from 0.04 m s^{-1} to 0.06 m s^{-1}, which resulted in the floodplain producing 74% of the total average discharge. Based on field observations of flow through floodplain vegetation, the Manning's n was overestimating the flow velocity and was not used to produce the peatland inflow rating curve (Fig. C.3).

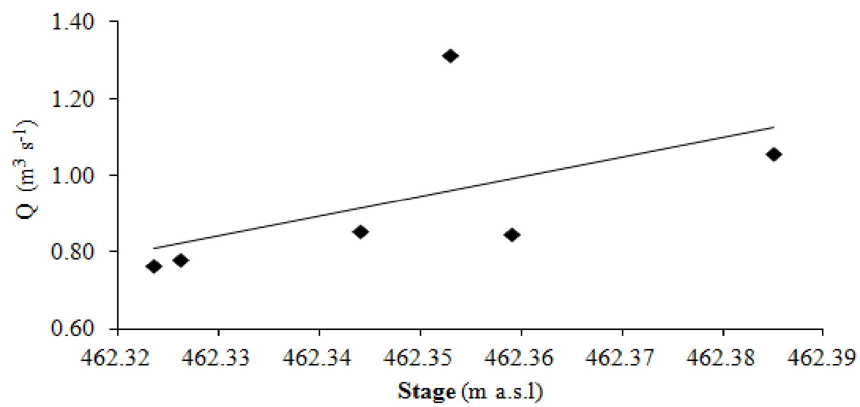


Figure C.3: Pine Fen Creek inflow rating curve using Manning's n for above bankfull stage (0.65 m or 462.37 m a.s.l.) with r^2 = 0.30, n = 6, P = 0.26 and Q = -2372.2 + 5.1327 • stage.

C.2 Peatland Outflow

Continuous stage recorded upstream of the culverts at the peatland outlet were compared with discrete stage measurements downstream (staff gauge) of the culverts during the study period (Fig. C.4). Data show stream stage was higher downstream of the culverts. There was a significant difference between the median stages (Mann-Whitney U = 39.0, n = 20, p = <0.001) likely due to flow attenuation by the beaver dam located 10 m downstream of the culvert (Westbrook et al., 2006). However, there is no significant difference between median discharges calculated using stage recorded upstream and downstream of the culverts (Mann-Whitney U = 1521.0, n = 20, p = 0.787). Therefore, upstream measurements of stage were used to produce the peatland outflow rating curve (Fig. C.5).

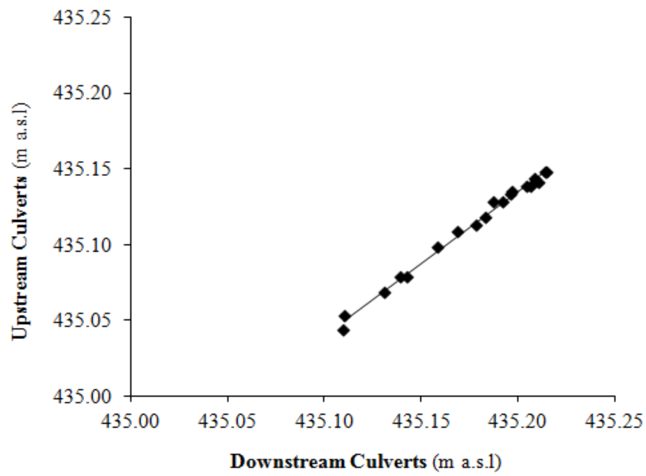


Figure C.4: Plot of stages recorded upstream (HOBO levelogger) compared with those measured downstream (staff gauge) of the culverts at Pine Fen outflow during the study period (r^2 = 0.99, n = 20, P = < 0.001).

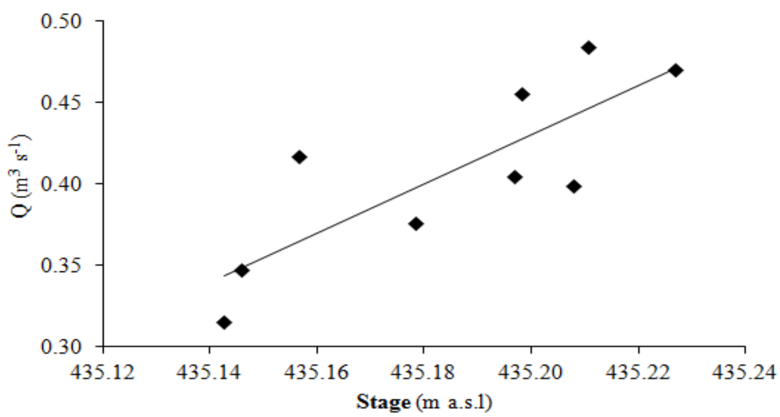


Figure C.5: Pine Fen Creek outflow rating curve for stages > 435.14 m a.s.l (level of east culvert sill = 434.51 m a.s.l.) with r^2 = 0.67, n = 9, P = 0.007 and Q = -657.2630 + 1.5112 • stage.

D.3 Ispuchaw Lake Outflow

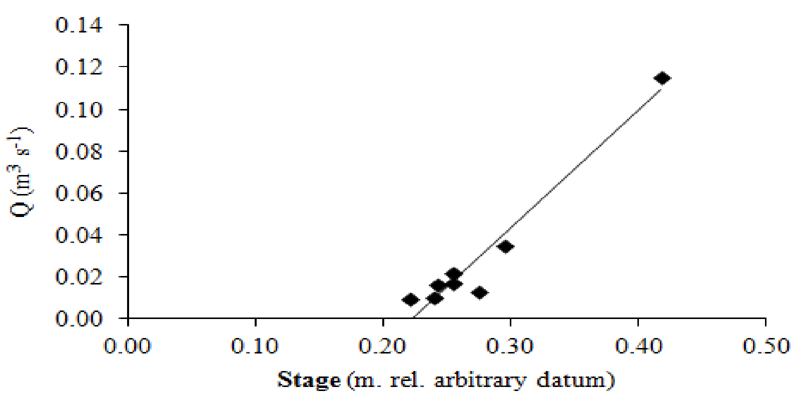


Figure C.6: Ispuchaw Lake outflow rating curve for stages > 0.22 m with r^2 = 0.95, n = 9, P = <0.001 and Q = -0.1247 + 0.5599 • stage.